# Bulletin
# of
# The Geological Society of America

VOLUME 59 September 1948 NUMBER 9

## CONTENTS

**Submarine geology of Bikini Atoll. By Kenneth O. Emery** ..... 855–860

**Reefs of Bikini, Marshall Islands. By J. I. Tracey, Jr., H. S. Ladd, and J. E. Hoffmeister** ..... 861–878

**Incandescent tuff flows in southern Peru. By C. N. Fenner** ..... 879–894

**Pleistocene climate and topography of the Arequipa region, Peru. By C. N. Fenner** ..... 895–918

**Further observations in the Ross Shelf Ice. By A. D. Howard** ..... 919–926

**High potash dikes in the Homestake Mine, Lead, South Dakota. By James A. Noble** ..... 927–940

**Stratigraphy and metamorphism in a part of the Black Hills and the Homestake Mine, Lead, South Dakota. By James A. Noble and J. O. Harder** 941–976

**Subscription $10 per year.**

**Publication Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.**

**Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York 27, N. Y.**

**NOTICE.—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.**

---

Entered as second-class matter in the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 3, 1918.

## PAPERS IN PRESS FOR FORTHCOMING ISSUES

PETROLOGY AND STRUCTURE OF THE PERCY QUADRANGLE. By R. W. Chapman. This paper is accompanied by a geologic map of the quadrangle, in color. Copies of the map (unfolded, in tubes) may be secured by sending remittance to The Geological Society of America, 419 West 117 Street, N. Y. 27, N. Y. Price 25 cents apiece.

GEOLOGY AND GROUND-WATER RESOURCES OF IWO JIMA. By Frank A. Swenson.

PETROGRAPHY OF IWO JIMA. By Gordon A. Macdonald.

GEOLOGY OF THE CONFUSION RANGE, WEST-CENTRAL UTAH. By C. S. Bacon, Jr.

KEY PERMIAN SECTION, CONFUSION RANGE, UTAH. By Norman D. Newell.

ELEVATED PHOSPHATE ISLANDS IN MICRONESIA. By L. E. Nugent.

OBSERVATIONS OF RELATIVE GRAVITY AT PARICUTIN VOLCANO, MEXICO. By Virgil E. Barnes and Frederick Romberg.

PORTION OF THE SAN ANDREAS RIFT IN SOUTHERN CALIFORNIA. By Robert E. Wallace.

GEOLOGY OF THE PALEOZOIC ROCKS, LOGAN QUADRANGLE, UTAH. By J. Stewart Williams.

AGE OF THE KINGSBURY CONGLOMERATE IS EOCENE. By Roland W. Brown.

Copies of Plate 3 of the Emery paper in the present issue may be secured by sending remittance to The Geological Society of America, 419 West 117 Street, N. Y. 27, N. Y. Price 25 cents apiece. Maps are unfolded and in tubes.

pies
log-
ece.

Virgil E.

allace.
Villiams.

sending
Y. Price

40′ 50′ 165°00′ 10′

12′ 00′

50′

40′

11′ 30′

20′

1000

2000

1000

1000

2000

BIKI

Ba

40′ 50′ 165°00′ 10′

BIKINI ATOLL — PRELIMINARY CHART

65° 00′ 10′ 20′ 30′ 40′ 50′

12° 00′ 50′ 40′ 11° 30′ 20′

1000 2000

BIKINI ATOLL

CONTOUR INTERVAL 200 FATHOMS

DOTS INDICATE POSITION OF SOUNDINGS

0 1 2 3 4 5 NAUTICAL MILES

SCALE

SOUNDINGS MOSTLY BY U.S.S. SYLVANIA

# ATOLL – PRELIMINARY CHART SHOWING OUTER SLOPES

Abstrac
Introdu
Outer s
Seamou
Lagoon
Bil
En
Bottom
Conclus
Referen

Plate
1. Bi
2. Ty
3. En
4. Bo
5. Co

The
the su
tion o
Atoll
Betwe
at the
side by
slopes
below
has a
side of
the lag
middle
covere
age of
to fine

A nu
(Opera
tion of

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 855-860, 5 PLS. SEPTEMBER 1948

# SUBMARINE GEOLOGY OF BIKINI ATOLL

BY K. O. EMERY

## CONTENTS

| | Page |
|---|---|
| Abstract | 855 |
| Introduction | 855 |
| Outer slopes | 856 |
| Seamounts and deep "terraces" | 857 |
| Lagoons | 857 |
| Bikini | 857 |
| Eniwetok | 857 |
| Bottom materials | 858 |
| Conclusions | 859 |
| Reference cited | 859 |

## ILLUSTRATIONS

| Plate | Facing page |
|---|---|
| 1. Bikini Atoll—Preliminary chart showing outer slopes | 855 |
| 2. Typical echo-sounding recordings in Bikini lagoon | 856 |
| 3. Eniwetok Atoll—Contoured chart of lagoon | 858 |
| 4. Bottom materials of Bikini lagoon | 860 |
| 5. Corals of Bikini lagoon | 861 |

## ABSTRACT

The program of investigations made in connection with the atomic bomb tests in the summer of 1946 at Bikini Atoll provided a rare opportunity for obtaining information on the physiography and sediments of atolls. The outer slopes around Bikini Atoll rise from a depth of about 2500 fathoms in a long gradually steepening curve. Between 0 and 200 fathoms the average slope is 25°. On the windward side, the reef at the top of the slope is bordered by a narrow 15-fathom terrace, but on the leeward side by a steep, locally vertical, cliff extending to 35 fathoms. Contours of the outer slopes reveal that a flat-topped mass having nearly the area of Bikini lagoon but below 700 fathoms extends to the northwest. The lagoon itself is saucer-shaped and has a maximum depth of 34 fathoms. Around the lagoon and bordering the inner side of the reef and islands is a 12-fathom terrace. Coral knolls rise abruptly from the lagoon floor, and some of them reach to within a few fathoms of the surface. The middle of the lagoon is floored chiefly by algal debris, whereas the shallow edges are covered by algal and foraminiferal sand. Coral comprises a relatively small percentage of the bottom materials. The outer slopes consist of algal debris grading outward to fine sand and finally to *Globigerina* sand.

## INTRODUCTION

A number of interesting scientific results were obtained from the atomic bomb tests (Operation Crossroads) at Bikini Atoll in the summer of 1946, aside from the evaluation of the bomb as a weapon. The major effort of the expedition was necessarily

bent toward studies likely to be important in connection with the bomb, but the number of ships available and the staff made up of specialists in many fields of science virtually insured the making of new discoveries. Submarine geology is one of the fields most likely to gain as a result of the trip, simply because the underwater portions of atolls had not been well studied previously. Modern sampling methods, recording echo sounders, and the underwater camera served as new tools for collecting data. To determine what changes in the topography and sediments were produced by the bomb, it was essential first to understand clearly the character of the bottom before the bomb blast, not only near the bomb site but also some distance away in the lagoon. Detailed studies were also made of the outer slopes, because of the possibility that landslides might be triggered by the explosion. Finally, to determine whether Bikini is a typical atoll and to provide controls for estimating bomb damage to organisms, the near-by Eniwetok, Rongelap, Rongerik, and Ailinginae atolls were studied briefly.

The geological investigations were supported by Military Intelligence Division, Office of the Chief of Engineers, and by the Bureau of Ships and formed part of the work of the Oceanographic Section of Joint Task Force One, under the leadership of Comdr. Roger Revelle.

## OUTER SLOPES

Bikini is one of several atolls at the north end of the Marshall group. The region is in the trade-wind belt, with winds blowing from the east and northeast. The water current and the swell come from the same general directions. In this part of the ocean most of the sea floor lies at a depth of 2500 to 3000 fathoms, or 15,000 to 18,000 feet. The foundations of the coral reefs rise on long gradually steepening slopes above the general bottom level.

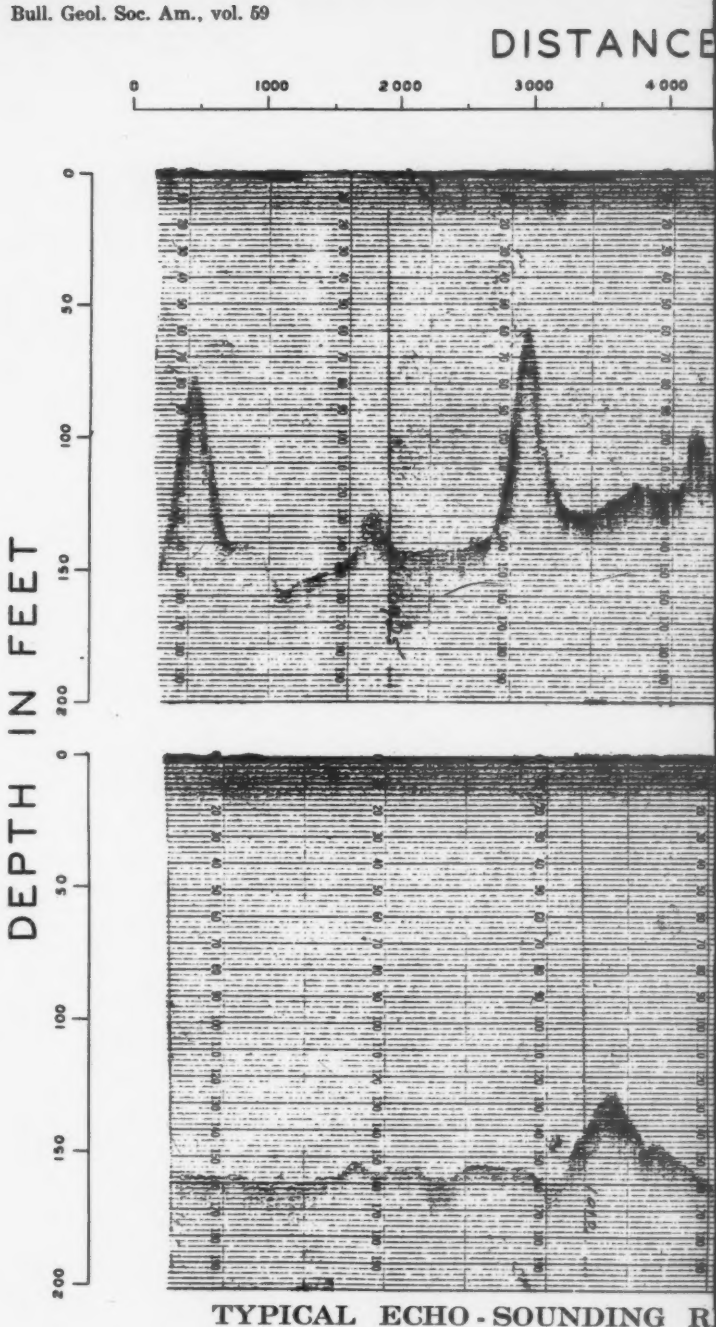
Plate 1 shows a preliminary chart of the outer slopes of Bikini Atoll with a 200-fathom contour interval. Dots indicate the position of the soundings used in drawing the contours; however, all the soundings were taken by a recording echo-sounding device which makes a continuous profile from which soundings can be read at any convenient spacing. About 8000 miles of such sounding records were made on the outer slopes and in the lagoons of the atolls. The slopes shown by the chart are gentle near the base and progressively steeper nearer sea level. Between 0 and 200 fathoms the average slope is 25°. The smoothness of the outer slopes is interrupted by spurs which lie seaward of the major projections and angles of the reef and extend from the surface down to the base of the atoll.

Between the reef projections and within the indentations on the east and north sides of the atoll there is a 10- to 15-fathom terrace. Its outer edge is indicated by a dashed line in Plate 1. On the opposite side of the atoll, the leeward side, the shallow terrace is absent, and in its place a steep, locally vertical, slope borders the reef edge and extends to a depth of about 35 fathoms.

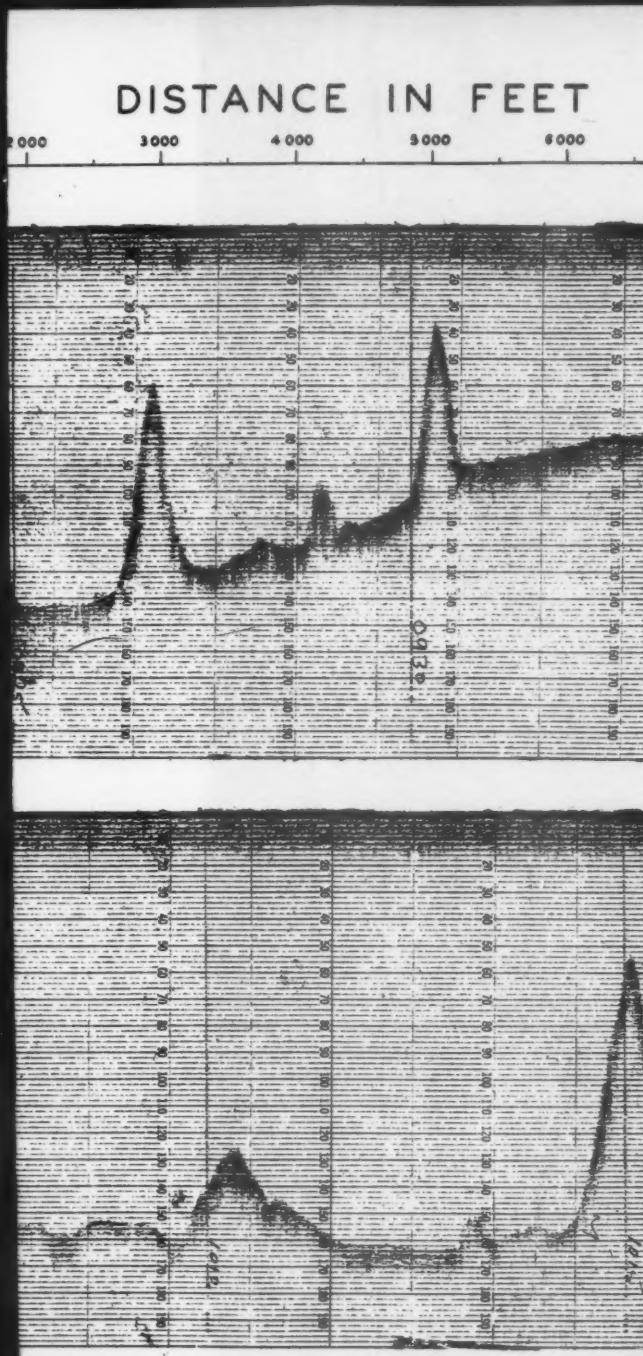
The outer slopes of the other atolls surveyed are generally similar in steepness and in the presence of spurs and shallow terraces. Rongerik, and Ailinginae atolls lie so close to each other that their slopes intersect at a depth of about 1000 fathoms.

1951
1981
UM

the
nce
the
or-
ds,
ng
ced
om
the
si-
ine
ge
re
on,
he
of
ou
er
he
20
es
0-
y
d-
at
on
re
00
dd
id
th
a
w
e
id
o



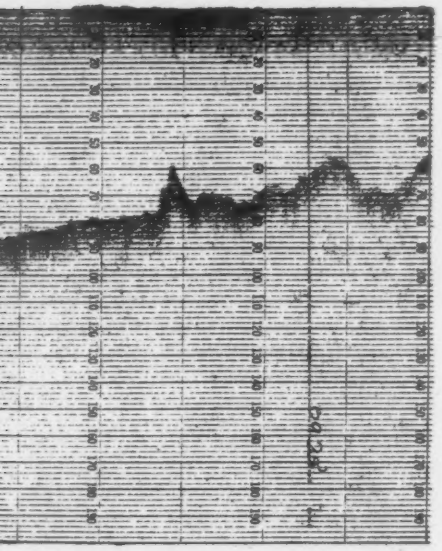

TYPICAL ECHO-SOUNDING R

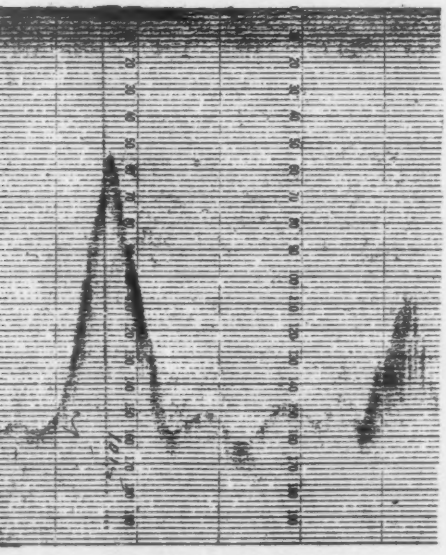



CHO-SOUNDING RECORDINGS IN BI

ET

6000 7000 8000 9000





S IN BIKINI LAGOON

## SEAMOUNTS AND DEEP "TERRACES"

One of the most interesting features shown by Plate 1 is a seamount which is tied to the northwest side of Bikini Atoll by a narrow deep neck. The seamount has a minimum depth of 680 fathoms, and its outer edge lies at about 750 fathoms; accordingly, it is extremely flat-topped. The area of the flat top is approximately equal to that of Bikini lagoon. Surveys were also made of several other seamounts which are more isolated. The tops of all the seamounts are flat and lie between depths of 550 and 1100 fathoms.

Deep terracelike structures that closely resemble the seamount tied to Bikini fringe some of the slopes of Eniwetok Atoll at about 750 fathoms. One lies south of the atoll, another northwest of it, but there is no indication of similar structures on the northeast and the southwest sides.

The flatness of the seamounts and the terracelike structures may be the result of past wave erosion, but heavy dredging equipment was not available for checking this possibility. Hess (1946) has described the Eniwetok structures and similar flat-topped seamounts and has termed them guyots.

## LAGOONS

### BIKINI

In contrast to the outer slopes, the lagoons of the atolls are very complicated and have far more detail than can be shown on small-scale charts. A chart of Bikini lagoon based only on Japanese lead-line soundings shows that the lagoon is a broad saucer having a 10-fathom terrace near the reef separated by a fairly steep slope from the middle deeper area. Numerous coral patches, or coral knolls, are present, and many of them rise to within a few fathoms of the surface. Much finer detail will be available when all the expedition's soundings have been plotted. Preliminary plotting of part of the soundings shows that a large depression in the 10-fathom terrace extends about 8 fathoms below the general level of the terrace.

The soundings reveal many coral knolls, and they appear to be slightly more abundant near the slope between the terrace and deeper part of the lagoon. Two typical lagoon profiles drawn by a recording echo sounder are shown in Plate 2. One of the profiles was made at right angles to the slope from the 10-fathom terrace to the deeper water of the lagoon, and the other is typical of the deep lagoon area. Both profiles show that the coral knolls rise abruptly from an otherwise fairly smooth floor. Several coral knolls were surveyed in detail with lead lines or were studied in diving operation. Some are very steep, nearly vertical, while others consist mostly of dead and loose coral fragments which form a kind of talus slope of about 45°. During the echo-sounding work it was discovered that the sounding records made over hard sandy bottom were much denser and narrower than those over the soft algal debris near the center of the lagoon; therefore, some interpretation of bottom composition can be made from the acoustic properties of the bottom.

### ENIWETOK

The lagoon at Eniwetok Atoll has the same kind of topography as Bikini lagoon. About 200,000 soundings from recording echo-sounding tapes were plotted in pre-

liminary form by the Navy prior to the expedition. These soundings served as a basis for the contour chart of Plate 3. About 2300 coral knolls are shown within the lagoon, giving it an irregularity which departs markedly from some of the conventional ideas of billiard table flatness of coral lagoon floors. Many of the coral knolls are so steep that the drawing of contours around them was impractical, so these are indicated by hachures.

A 10-fathom terrace borders the lagoon side of the reef along the east, north, and northwest sides of the lagoon. The terrace is widest where the reef bends outward away from the lagoon and narrowest where the reef is indented toward the lagoon. This relationship also exists at Bikini Atoll where, in addition, surveys of the outer slopes showed a shallow terrace seaward of the reef indentations. In the northwest part of the lagoon, where the terrace is widest it contains a depression which extends about 8 fathoms below the terrace surface, like the one at Bikini. The physiographic relationships suggest that the 10-fathom terrace was once a reef surface, upon which the present, younger reef has grown.

A deep channel drains eastward through one of the passes in the reef (Pl. 3). At the point where the channel crosses the reef and the 10-fathom terrace its depth is about the same as the maximum depth of the lagoon. Each atoll has one such channel: Bikini at the southwest, Rongelap at the southeast, and Rongerik at the south side. The channel of each atoll becomes shallower toward the center of the lagoon; the channels may have originated as outlets of lakes left by a glacially lowered sea level.

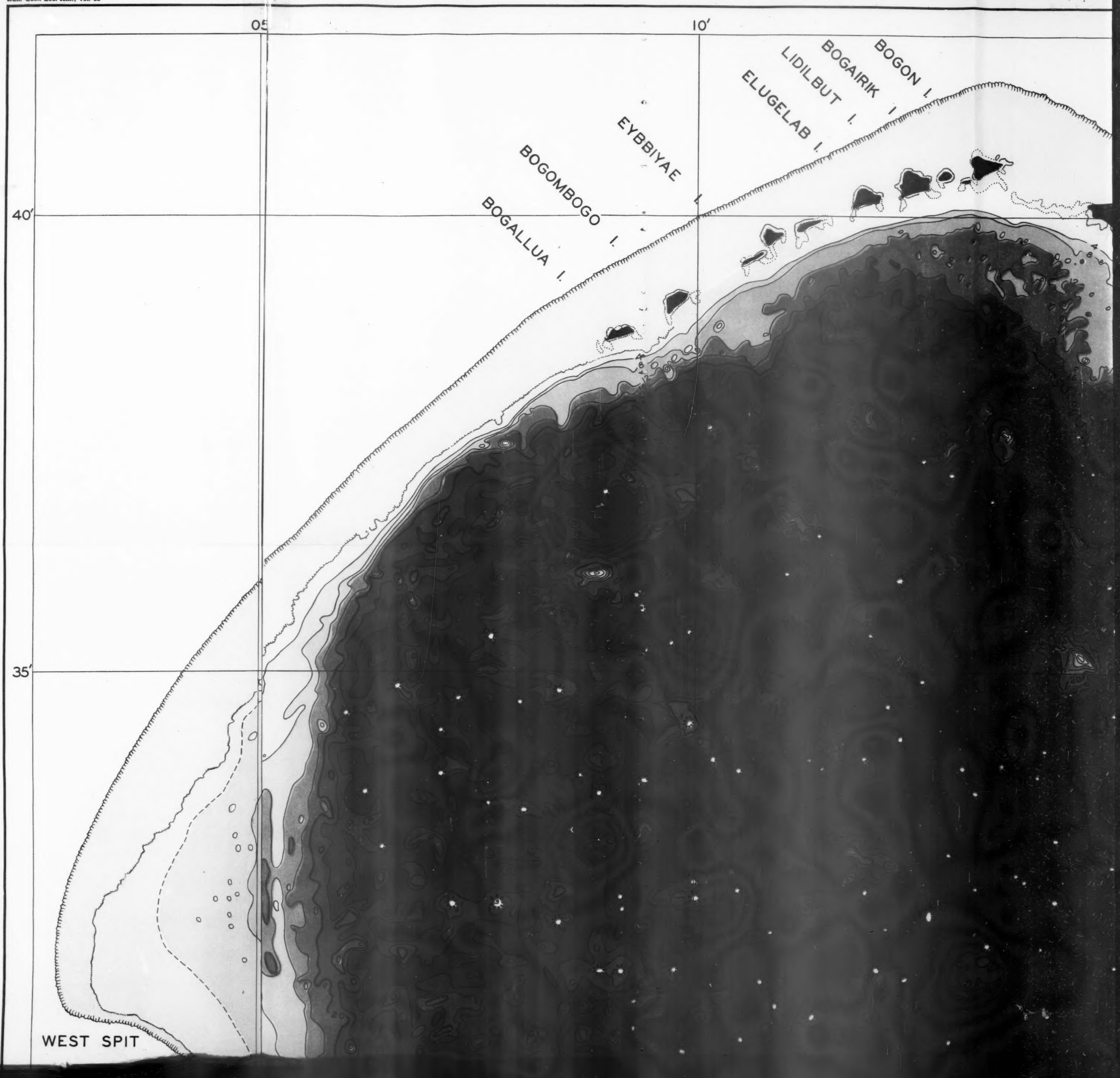
## BOTTOM MATERIALS

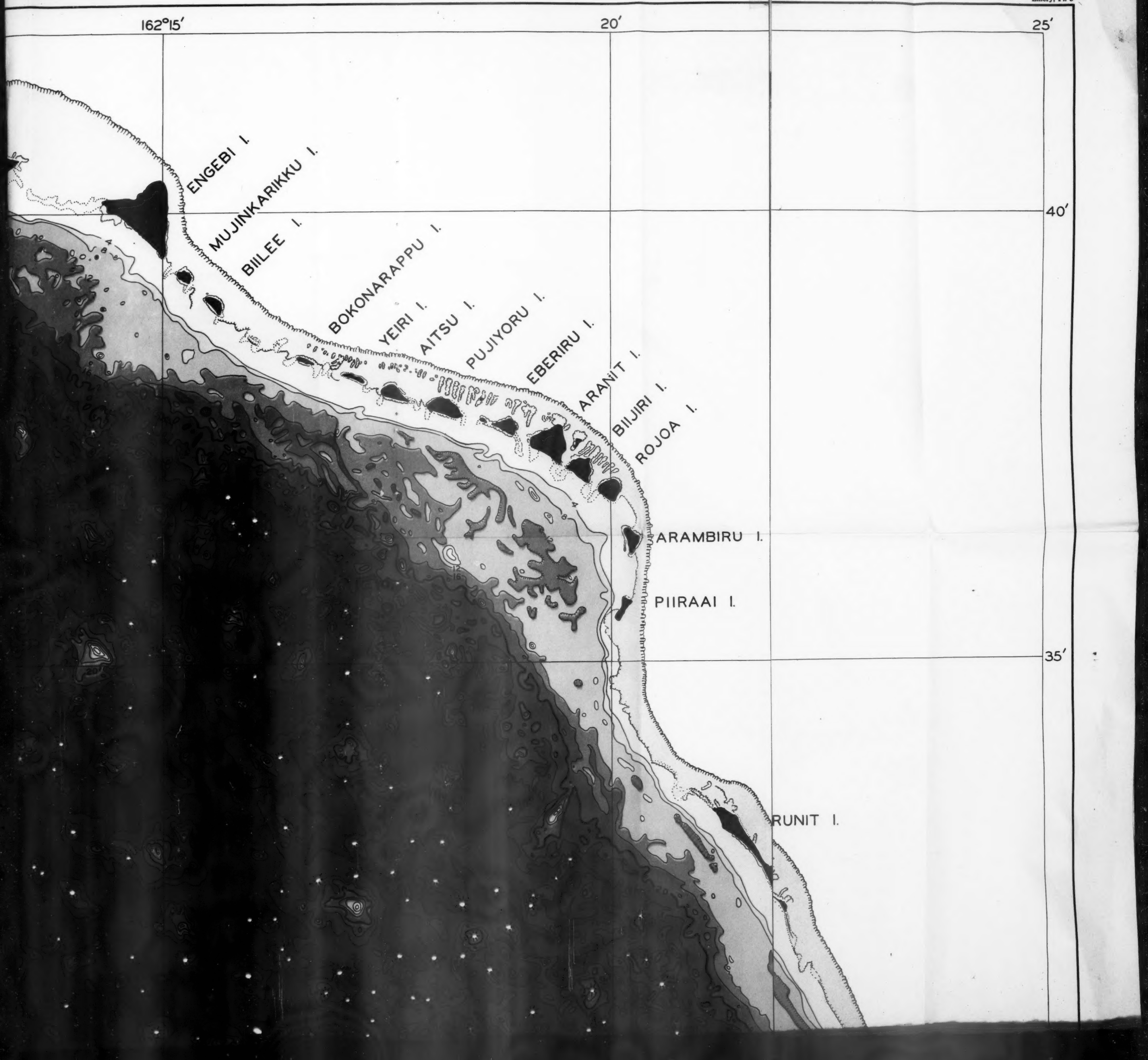
There are several types of sediments or bottom materials in the submarine portions of the atolls. The collection of about 2500 bottom samples permits the making of fairly detailed sediment-distribution charts, both inside the lagoons and around the atolls. Nearest shore in Bikini lagoon is a coarse foraminiferal sand, a continuation of the beach sand. This grades lagoonward into a narrow zone of fine sand. At about 20 fathoms the sand gives way to *Halimeda* debris which covers most of the lagoon floor. *Halimeda*, a kind of calcareous alga, appears to be the main contributor to the sediments of the atolls. When alive, it is a string of oval, flat, green segments which break apart and bleach to white when the plant dies (Pl. 4, fig. 1). Branching corals are present in many samples, but the chief areas of coral growth are the coral knolls.

During the course of the work about 700 underwater photographs were taken, mostly in the lagoons. They show that *Halimeda* in depths of about 30 fathoms is a thin carpet of living plants over an unknown thickness of white loose *Halimeda* debris. Nearer shore, various algal and coral fragments lie on a sandy floor (Pl. 4 fig. 2). Corals found on the shallow reef and the shallow tops of coral knolls are mostly fairly massive or bracketlike (Pl. 5, fig. 1). At 10 or 15 fathoms, however a branching form dominates (Pl. 5, fig. 2).

The sediments of the outer slopes of the atolls appear to be somewhat simpler in their distribution. Beyond the reef edge, the bottom deepens very steeply, especi-

V591
1948
UM


05'
10'
40'
35'
BOGON I.
BOGAIRIK I.
LIDILBUT I.
ELUGELAB I.
EYBBIYAE I.
BOGOMBOGO I.
BOGALLUA I.
WEST SPIT


162°15′
20′
25′
40′
35′
ENGEBI I.
MUJINKARIKKU I.
BIILEE I.
BOKONARAPPU I.
YEIRI I.
AITSU I.
PUJIYORU I.
EBERIRU I.
ARANIT I.
BIIJIRI I.
ROJOA I.
ARAMBIRU I.
PIIRAAI I.
RUNIT I.

WEST SPIT

11°
30′

25′

20′

0

GILI I.

GAINEM I.

LIBIRON I.

BOGAN I.

BUGANEGAN I.

IGURIN I.

10′

162°15′

SMALL STEEP CORAL KNOLLS
DEPTH TO TOP OF CORAL KNOLL IN FATHOMS

OUTER EDGE OF REEF

INNER EDGE OF REEF

SAND

# ENIWETOK ATOLL - LAGOON

1 0 1 2 3 4 5

STATUTE MILES

CONTOUR INTERVAL - 4 FATHOMS

1948

# ENIWETOK ATOLL-CONTOURED CHART OF LAGOON

CHINIEERO I.

JAPTAN I.

CHINIMI I.

BOGEN I.

MUTI I.

PARRY I.

ENIWETOK I.

11° 30′

25′

20′

162°15′

20′

25′

- LAGOON

3 4 5

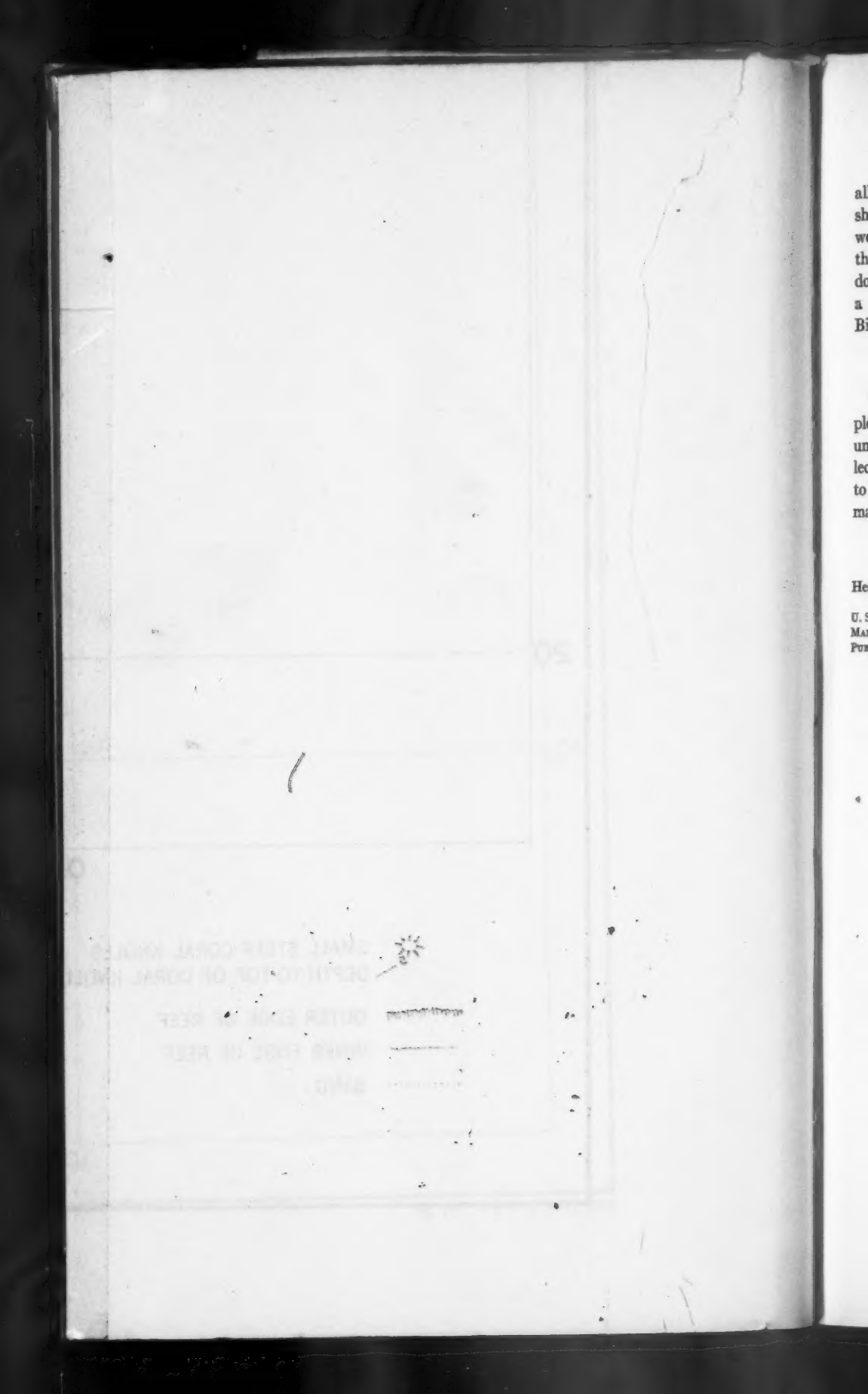
ES
FATHOMS

DEPTH ZONES - IN FATHOMS

32 28 24 20 16 12 8 4 0 REEF LAND

BASED ON BOAT SHEETS CONTAINING 180,000 SOUNDINGS
BY U.S.S. BOWDITCH - 1944

# CHART OF LAGOON



ally on the leeward side of the atoll. Probably the chief constituents of the steep shallow slope are corals and *Lithothamnion*. Beyond the steepest portion, corals were found to depths of only about 35 fathoms. *Lithothamnion* was found only in the same shallow depths. Between 20 and 200 fathoms, *Halimeda* debris appears to dominate. It grades through coarse sand to fine sand at about 500 fathoms. Half a dozen short core samples from the flat surface of the deep seamount bordering Bikini Atoll consisted only of *Globigerina* sand.

## CONCLUSIONS

The charts and the data outlined are preliminary and are based chiefly on field plotting and field estimates. They should not be used for drawing many conclusions until they are verified and supplemented by more detailed studies of the large collections of bottom samples and soundings. Several years will probably be required to complete the study of all the data, but it is believed that the results of the study may throw much light on the various theories of reef origin.

## REFERENCE CITED

Hess, H. H. (1946) *Drowned ancient islands of the Pacific Basin*, Am. Jour. Sci., vol. 244, p. 772–791.

U. S. Geological Survey, Washington 25, D. C.
Manuscript Received by the Secretary of the Society, September 3, 1947
Published by Permission of the Director, U. S. Geological Survey

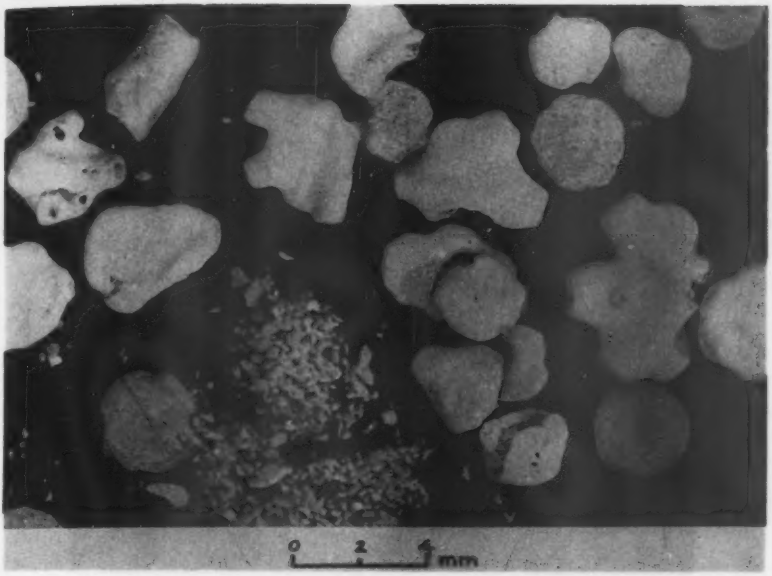



FIGURE 1. *Halimeda* DEBRIS
Microphotograph shows loose segments of *Halimeda* with associated Foraminifera and fine-grained sediment.

FIGURE 2. SAND WITH SCATTERED CORALS AND ALGAE
Underwater photograph (No. 77) in 9 fathoms. Camera is inclined 45°. Length of bottom edge is about 6 feet.

FIGURE 1. CORAL KNOLL
Underwater photograph (No. 585) in 10 fathoms. Camera is horizontal. Length of bottom edge is about 5 feet.

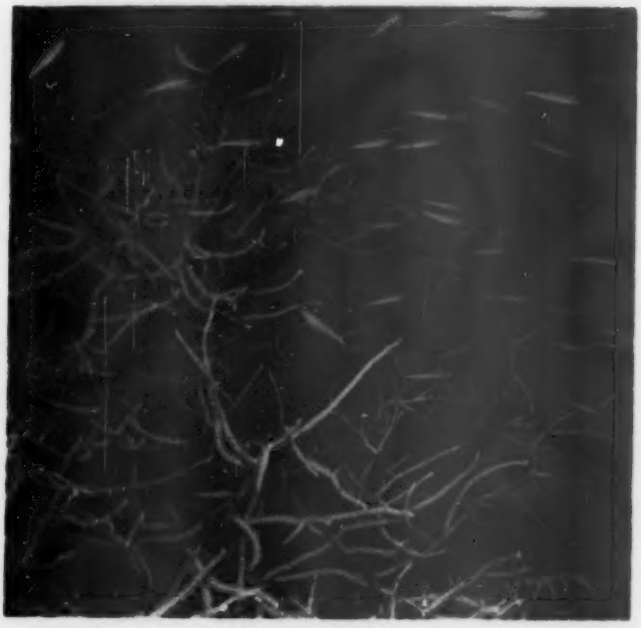


FIGURE 2. BRANCHING CORAL
Underwater photograph (No. 88) in 10 fathoms. Note the school of fish above the coral. Length of bottom edge is about 4 feet.

FIGURE 1. *Lithothamnion* RIDGE, BIKINI ISLAND

View showing algal buttresses and surge channels. Note open pool on reef, lower left foreground. Reef type II-B-(1).

FIGURE 2. GROWTH OF ALGAE (PURPLISH-RED), HYDROZOANS (YELLOWISH-GREEN PLATY *Millepora*) AND CORAL (BRANCHING *Acropora*) NEAR EDGE OF REEF

Growth of this type is especially characteristic of lee reefs adjacent to passes.

*LITHOTHAMNION* RIDGE AND GROWTH OF CORALS AND ALGAE NEAR REEF EDGE

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 861-878, 8 FIGS., 11 PLS. SEPTEMBER 1948

# REEFS OF BIKINI, MARSHALL ISLANDS

BY J. I. TRACEY, JR., H. S. LADD, AND J. E. HOFFMEISTER

## CONTENTS

| | Page |
|---|---|
| Abstract | 862 |
| Introduction | 862 |
| Description of reefs | 863 |
| Foundation of the existing reef | 863 |
| Reef zonation | 863 |
| General discription | 863 |
| Marginal zone | 867 |
| Coral-algal zone | 867 |
| Reef flat | 867 |
| *Heliopora* zone | 867 |
| Beach-rock and beach | 868 |
| Lagoon reefs | 868 |
| Classification of main reef types | 868 |
| Key | 868 |
| Distinctive features | 870 |
| Development of the marginal zone | 872 |
| Offshore algal spurs and grooves | 872 |
| Surge channels | 874 |
| General statement | 874 |
| Room-and-pillar structure | 874 |
| Blowholes | 874 |
| Erosion of the marginal zone | 876 |
| Description | 876 |
| Re-entrants formed by collapse | 876 |
| Comb structures suggesting progressive wave erosion | 876 |
| Summary | 877 |

## ILLUSTRATIONS

| Figure | Page |
|---|---|
| 1. Sketch map showing the distribution of terrace in lagoon and outside reef | 864 |
| 2. Generalized cross section through the windward reef of Bikini | 865 |
| 3. Sketch map showing the distribution of main reef types | 869 |
| 4. Cross section through marginal zone and algal spurs | 873 |
| 5. Sea reef off Tufa Island, Rongelap | 873 |
| 6. Diagram illustrating the development of room-and-pillar structure | 875 |
| 7. Erosional re-entrants on southern reefs | 877 |
| 8. Lee reef off Burok Island, Rongelap | 877 |

| Plate | Facing page |
|---|---|
| 1. *Lithothamnion* ridge and growth of corals and algae near reef edge | 861 |
| 2. Reef and submarine terrace | 864 |
| 3. Grooved reef with submarine terrace. Eroded reef | 865 |

| Plate | | Facing page |
|---|---|---|
| 4. | Reef zonation | 866 |
| 5. | Ungrooved margin of sea reef | 867 |
| 6. | Coral-algal zone | 868 |
| 7. | Coral growth and *Heliopora* zone | 869 |
| 8. | Shore features and blowholes | 872 |
| 9. | Marginal grooves and room-and-pillar structure | 873 |
| 10. | Erosion on southern reefs | 876 |
| 11. | Reef cavern and pools | 877 |



## ABSTRACT

Systematic surveys made at Bikini before and after the atomic bomb tests Operation Crossroads) in the summer of 1946 afforded an unusual opportunity to examine the reefs and islands of one of the larger atolls of Micronesia. The existing reef appears to be developed on an older surface that extends as a terrace into the lagoon and beyond the seaward margin of the reef. A sketch map shows the distribution of the terraces in relation to the existing reefs.

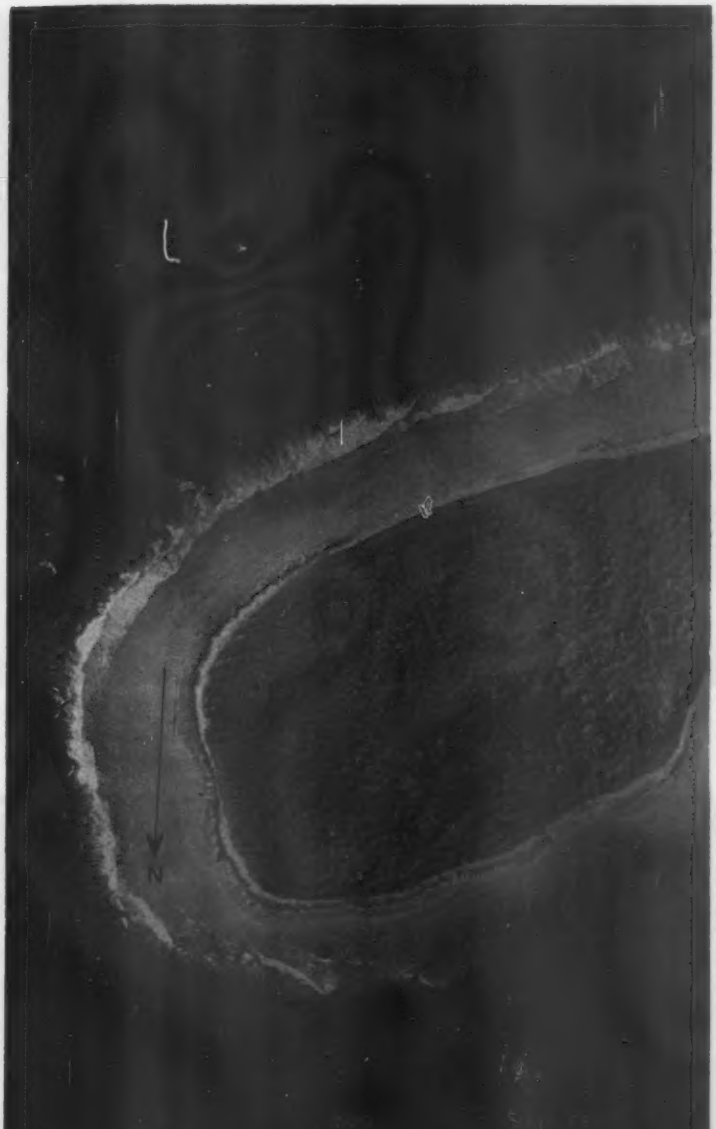
Reef zonation is briefly described, and a classification of reef types based upon the characteristics of the marginal zone is presented. Illustrations are given from Bikini's reefs and those of near-by atolls. A sketch map shows the distribution of the reef types at Bikini relative to prevailing winds and currents. The processes of reef growth and unusual areas wherein the existing reef appears to be undergoing destruction are also described and figured.


## INTRODUCTION

Studies of the geology of Bikini and several near-by atolls in the Marshall Islands group were made in connection with Operation Crossroads carried out in 1946 by Joint Task Force One. The geological investigations were supported by Military Intelligence Division, Office of the Chief of Engineers, and by the Bureau of Ships, and formed part of the work of the Oceanographic Section of Joint Task Force One under the leadership of Comdr. Roger Revelle. The operation afforded an unusual opportunity to examine in detail the reefs and islands of several of the larger atolls. A comprehensive report is now in preparation. The present paper is limited in scope; it classifies and describes briefly the types of reefs found and interprets the later stages of reef growth. Speculation as to the origin of the atoll is withheld pending the completion of additional studies planned for the area. It appears unlikely that the conclusions here presented as to the later stages of reef growth will be greatly modified by additional investigations.

Prior to the atomic bomb tests, Tracey and Ladd studied the reefs and islands of Bikini for approximately two months. Comparative studies of Eniwetok, Rongerik, and Rongelap atolls were later made by Tracey and Charles Bates of the Woods Hole Oceanographic Institution, now with the Division of Oceanography, U. S. N. Hydrographic Office. Following the bomb tests Hoffmeister joined the party which spent several weeks in a re-examination of many of the reefs and islands of Bikini.

At 26 localities widely spaced around Bikini, detailed traverses were measured across the reefs at right angles to the reef front. The traverses record observations along a strip approximately 20 feet in width; they show the relative abundance of




REEF AND SUBMARINE TERRACE

Seaward terrace off Arji Island, Bikini, adjacent to Enyu Channel. Airplane photograph shows margin of broad terrace near top of photograph; irregular "combtooth" structure of reef margin (type II-B-(2) ), center, and lagoon reef sloping to lagoon terrace at bottom of photograph. Airphoto by U. S. Navy.

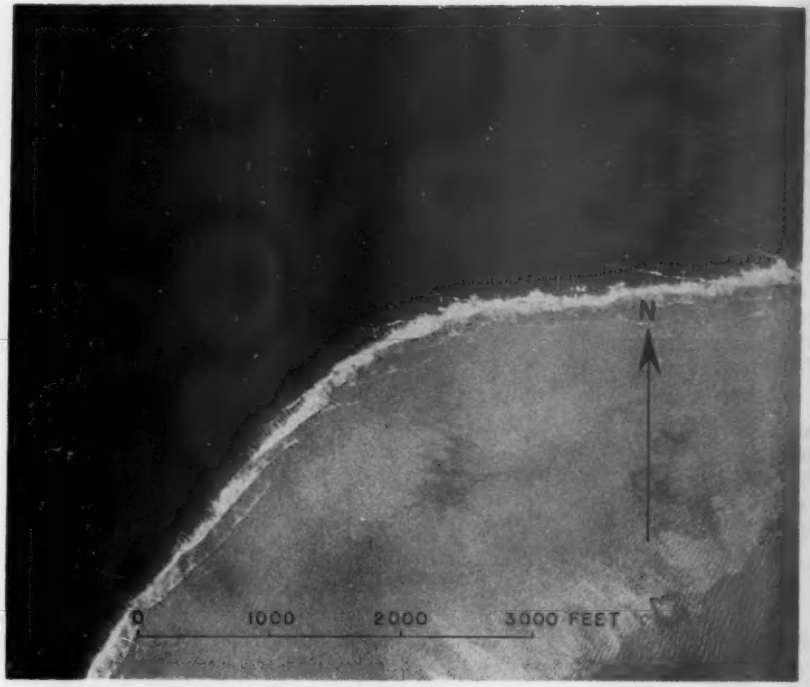



FIGURE 1. SEAWARD TERRACE, RONGERIK
Airplane photograph showing broad terrace outside reef margin on northwest part of Rongerik reef. The grooves in the marginal zone are well developed on the north edge but diminish in size and development on the northwest reef edge. (Reef type I-A). Airphoto by U. S. Navy.

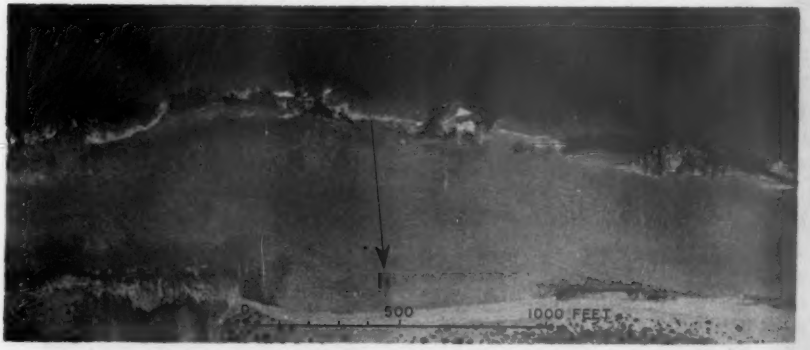



FIGURE 2. REEF TYPE II-B-(1)
Eroded reef off Eman Island. Aerial view of a southern reef showing slightly scalloped reef margin with large irregular erosional re-entrants. The marginal crest is the locus of greatest erosion. Airphoto by U. S. Navy.

GROOVED REEF WITH SUBMARINE TERRACE; ERODED REEF

corals and algae of the more important types and the distribution of channels, caverns, pools, and reef deposits. Observations of growth and erosion forms off the seaward margin of the reef and in the reef pools were made by swimming with a face mask. Similar studies in the lagoon were supplemented by diving and dredging. The excellent air photographs made by the Navy were of great value, particularly in studies involving zonation and distribution of reef types; a number are reproduced in the present paper. The remaining photographs (except Pl. 6, fig. 2) were taken by Tracey.

During Operation Crossroads a series of code names was adopted for many of the islets of Bikini. These short names are used in the present paper.

## DESCRIPTION OF REEFS

### FOUNDATION OF THE EXISTING REEF

At Bikini evidence suggests that the existing reef is growing on an older surface. In most places there is a terrace or floor that extends beyond the reef edge at a depth of 15–30 feet (generally about 20 feet), sloping seaward at an estimated 10°–15°. It may be seen in aerial photographs, but the nature of the surface is obscured by the water, and the apparent "outer edge" on the photograph is generally due to the limiting depth beyond which the seaward slope is not visible. Figure 1 shows the approximate distribution of this surface. It is absent along the west side of the atoll where the reef margin drops off steeply to 30 fathoms or more.

At many localities this surface is a terrace, with a main break in slope to 45° or more occurring at about 15 fathoms (Pl. 2).

Similar terraces are present on near-by atolls (Pl. 3, fig. 1). This older surface appears to extend beneath the existing reef, rising to within a few feet of the main reef flat. It apparently passes beneath the reef flat and islands, where they are present, and extends as a marginal terrace into the lagoon. The main part of the lagoon terrace lies at a depth of about 10 fathoms, its deeper edge reaching a depth of about 12 fathoms. The floor of the lagoon beyond the terrace reaches a maximum depth of about 35 fathoms. The inferred relation of the present reefs to this older surface is shown diagrammatically in Figure 2. Evidence for this relationship as observed on the seaward parts of the reefs will be discussed in a later section.

The depths of both the seaward and the lagoon terrace are concordant, and these two terraces may be connected beneath the present reef by a flat surface at a depth of about 10 fathoms. Such a surface might be a former reef level, or a surface of truncation. Its presence could be verified only by drilling.

### REEF ZONATION

*General description.*—It is impossible in most places to traverse a reef without recognizing that the area is at least roughly zoned parallel to the reef front. Some of the bands are growth zones wherein certain organisms outnumber all others; elsewhere the bands are depositional flats or areas of solid rock pavement. Differences in the composition of the reef surface and in organic growth are also observable laterally—along lines parallel to the reef front—but these differences are less striking than

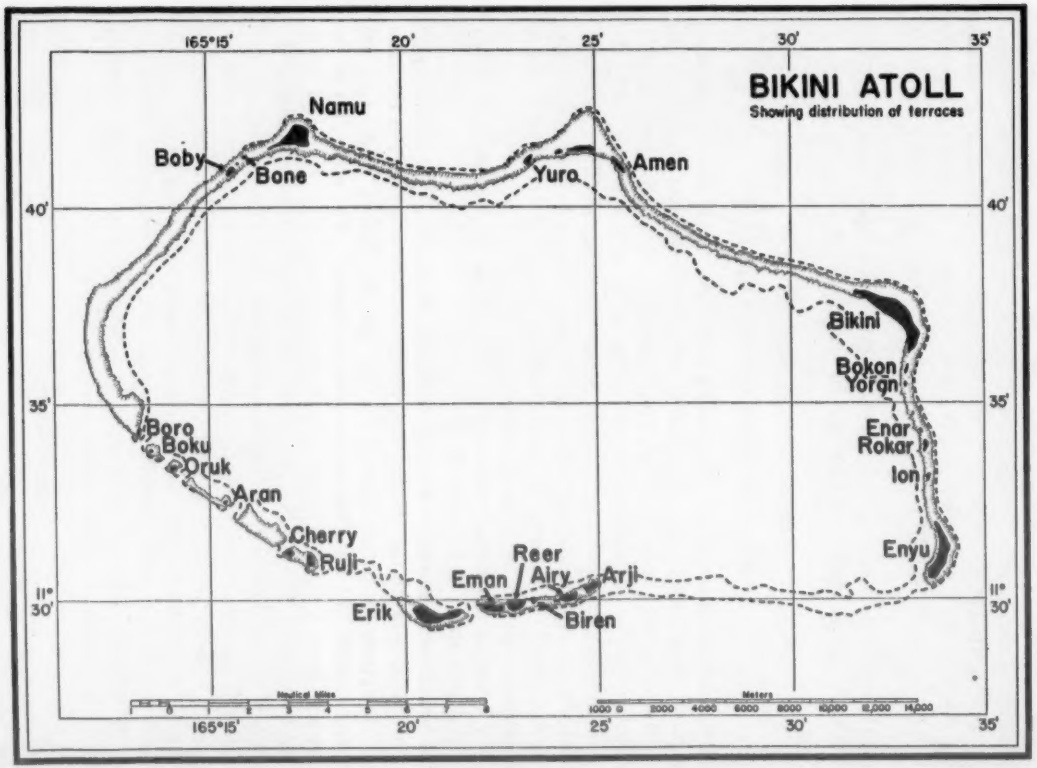



FIGURE 1.—*Sketch map showing the distribution of terrace inside lagoon and outside reef*

Seaward terrace generalized to include all areas with a gentle offshore slope. Terrace shown by heavy dashed line. Outline of reefs and islands from H. O. 6032.

FIGURE 1.—*Sketch map showing the distribution of terrace inside lagoon and outside reef*

Seaward terrace generalized to include all areas with a gentle offshore slope. Terrace shown by heavy dashed line. Outline of reefs and islands from H. O. 6032.

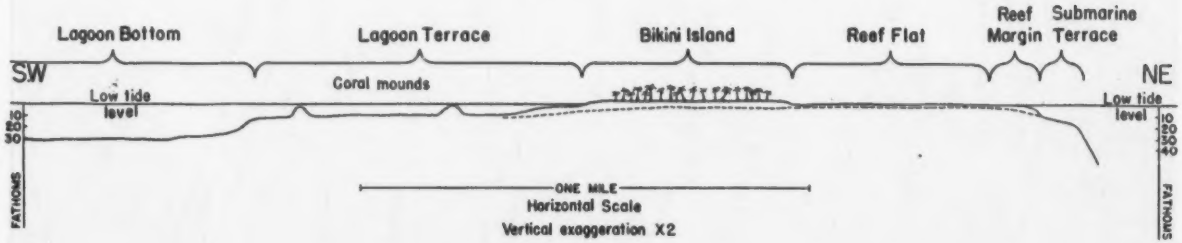
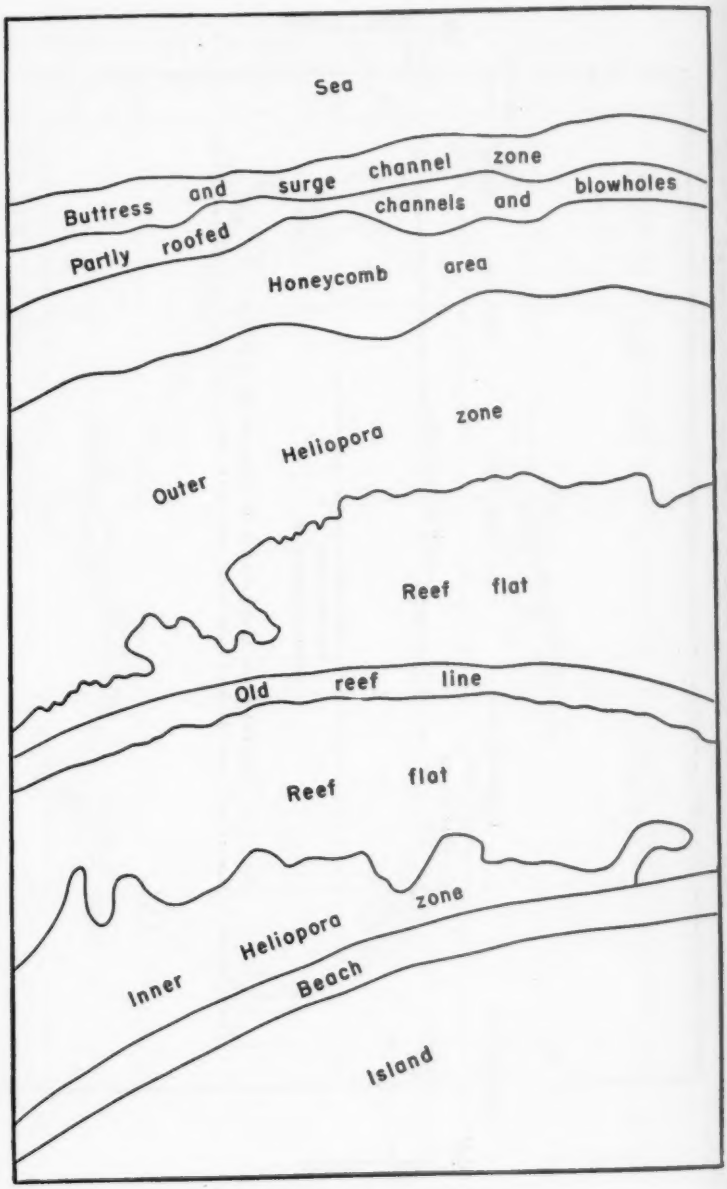



FIGURE 2.—*Generalized cross section through the windward reef of Bikini*

Figure shows relation of present reef, island, and lagoon bottom surface (solid line) to inferred older surface (dashed line) and terraces.




EXPLANATION OF PLATE 4

REEF ZONATION

Reef zonation. Airplane photograph of sea reef northeast of Bikini Island, showing division of the reef into zones parallel to the margin. Airphoto by U. S. Navy.

UNGROOVED MARGIN OF SEA REEF

**Reef type II-A.** Aerial view of the ungrooved, slightly scalloped reef edge of the broad leeward reef on the west side of Bikini Atoll. Zonation parallel to the edge is well developed. The irregular lagoon margin of the reef is in the middle ground; below is a shallow shelf 10 to 20 feet deep with a rich coral growth. Airphoto by U. S. Navy.

the banding. Zonation of both types is clearly shown on air photographs (Pls. 4, 5). The commoner types of zones from the seaward edge to the lagoon are briefly described below:

*Marginal zone.*—The seaward slope of the reefs on the windward side is cut by a series of closely spaced grooves. At the reef margin a colorful algal structure known as the *Lithothamnion* ridge is developed in the zone of the surf. In some places this is a low bulge, uncut by the grooves that are developed below low-tide level. In others, particularly in the reefs adjacent to islands, it is more prominent—a cuesta-like structure with a steep slope to the sea, cut by the landward extensions of the submarine grooves. These cuts through the ridge are called surge channels, and they divide the ridge into massive algal buttresses (Pl. 1, fig. 1).

The ridge is composed mainly of pink to dark-red *Lithothamnion*. Globular spongelike colonies of these algae cover the seaward sides of the buttreses and line the sides of the surge channel. On the gentle backslope of the ridge they give way to flat botryoidal types of "pavement" *Lithothamnion* that are usually somewhat lighter in color. Corals and hydroid corals occur on the ridge but generally cover less than 5 or 10 per cent of the surface area.

Leeward reefs in general are ungrooved and otherwise featureless structures (Pl. 5), The marginal zone of such a reef, though elevated but slightly above the main flat is the most interesting part of the reef. In some places it is the zone of richest organic growth (Pl. 1, fig. 2), in others the site of striking erosional features.

*Coral-algal zone.*—In most areas, the marginal zone is flanked on the landward side by a zone of rich coral growth 25 to several hundred feet wide (Pl. 6, fig. 1). In this zone true reef corals grow in greatest profusion and variety, in some areas covering more than half the reef surface (Pl. 6, fig. 2). The remainder of the zone in most places is covered by pavement and nodular types of pink *Lithothamnion*. The algae surround the heads of living coral and in many places seem to be encroaching upon them; elsewhere the large masses of coral—chiefly *Acropora* with short stumpy branches rising from a broad flat base—are encroaching upon the algae. The two form a solid mass. The zone is covered by only a few inches of water at lowest tide, but even then an occasional large wave wets the projecting parts of the coral heads. The coral-rich zone is much broader on reefs between islands than on reefs fringing islands. On leeward reefs the variety of corals in the zone exceeds that on windward reefs.

*Reef flat.*—This zone, in many places, forms the major part of the reef. On windward reefs it is generally a barren flat, composed typically of eroded coral and algal limestone covered with a thin veneer of sand, living Foraminifera, and soft brownish algae. Colonies of living coral are widely scattered, being in general more abundant to seaward. Hardy corals such as *Porites* and *Favia* outnumber other types. Most of the reef flat is covered by 6 inches to 1 foot of water at extreme low tide, but large areas near shore may be completely exposed except for small pools that fill shallow depressions in the surface.

On leeward reefs the flat is rougher, more irregular, and richer in coral growth than on windward reefs (Pl. 7, fig. 1). Hollows on the floor are filled with foraminiferal and coral sand; such areas increase in size and number landward or lagoonward.

*Heliopora zone.*—In many areas the surface of the reef flat is depressed and is

covered by 1–4 feet of water at low tide. On the reef floor large subcircular colonies of the alcyonarian, *Heliopora*, 3 to 25 feet or more across rise to low-water level. They are flat-topped, and the living part of the colony grows in radiating handlike forms around the circumference of the colony. As they grow, the heads coalesce to form a nearly continuous flat at low-tide level, broken only by the pools or passages between heads that have not yet grown together (Pl. 7, fig. 2). Such zones are well developed off parts of Bikini Island. In the area shown on Plate 4 both an inner and an outer *Heliopora* zone may be recognized.

*Beachrock and beach.*—Where islands are present, the reef flat may terminate against a zone of beach rock. This zone may be a belt or groin of lithified gravel or boulders 1 to 3 feet above low tide or, more rarely, a bedded beach sand cemented into rock (Pl. 8, fig. 1). Behind this zone there is generally a beach whose sand grains are mostly the shells of the Foraminifera that live in countless numbers on the rocky flat. Where beach rock is not developed the reef flat extends to the edge of the beach.

On many leeward islands, particularly those adjacent to passes, the beach is replaced by a rampart of loose coral head boulders (Pl. 8, fig. 2)

*Lagoon reefs.*—The margins of the reef that face the lagoon vary almost as widely as do the seaward margins. Along the windward side of the atoll where the lagoonal reef lies in the lee of the main reef flat its margin may be serrate in plan. The extensions of the reef into the lagoon commonly slope gently below low-tide level to depths of 5 to 10 feet. Beyond this line—which is very irregular—there is a steeper slope to the surface of the lagoonal terrace, the deeper edge of which lies at about 12 fathoms. Patches of living coral are scattered along the margin of the reef. Some are 100 feet wide and several hundred feet long and extend upward nearly to low-tide level.

In a few places the lagoonal edge of the reef on the windward side drops steeply to depths of 20–30 feet. Such a condition exists along the eastern side of Rongelap. The edge of the lagoonal reef is undercut, and large slabs of what was formerly the reef flat have been broken off and lie on the floor of the lagoon. There is little or no coral growth in such areas.

On the lee side of an atoll the lagoon reef is exposed to the full force of any waves generated in the lagoon by the prevailing wind. The growth of corals is much richer, and there are numerous patches 5 to 50 feet in diameter that rise close to the surface from depths of 5 to 20 feet. These areas are well separated, and in most places the edge of the reef is very irregular (Pl. 5, center). In a few places the coral areas coalesce, and *Lithothamnion* grows in sufficient abundance to form a marginal zone comparable to the margin of a seaward reef. Examples of this type of lagoonal reef are found on the north and west sides of atolls. The lagoonal reef off Namu Island at Bikini is a good example, but even more striking is the lagoonal reef in the northwest quadrant of Eniwetok.

## CLASSIFICATION OF MAIN REEF TYPES

### KEY

This classification of reefs is based primarily upon the characteristics of the marginal zone of the seaward side because this zone is the most vital part of the reef.

FIGURE 1. CORAL-ALGAL ZONE
A lee reef southwest of Erik Island. The reef floor is covered by one or two inches of water and the coral colonies are completely exposed at low tide. The coral is dominantly *Acropora*.

FIGURE 2. CORAL-ALGAL ZONE
Vertical view showing a rich growth of *Pocillopora* (center), nodular *Porites* (left) and several species of *Acropora*. Width of photograph about 2 feet. Photograph by H. S. Ladd.

CORAL-ALGAL ZONE

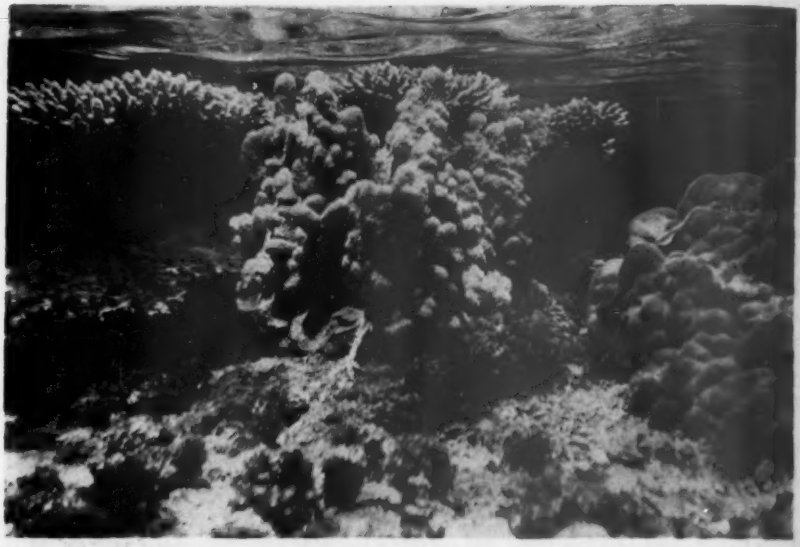


FIGURE 1. UNDERWATER PHOTOGRAPH SHOWING TWO SPECIES OF *Acropora* (CENTER) AND A MASSIVE *Porites* (RIGHT) GROWING ON THE REEF FLAT
Height of corals about one foot.

FIGURE 2. *Heliopora* ZONE
View of sea reef, Bikini Island, taken from beach. The area is the same shown in Plate 4. The *Heliopora* colonies grow radially and coalesce to form a reef platform at low tide level. The "old reef line" shown in Plate 4 appears in the middle distance; beyond, the surf is breaking on the *Lithothamnion* ridge.

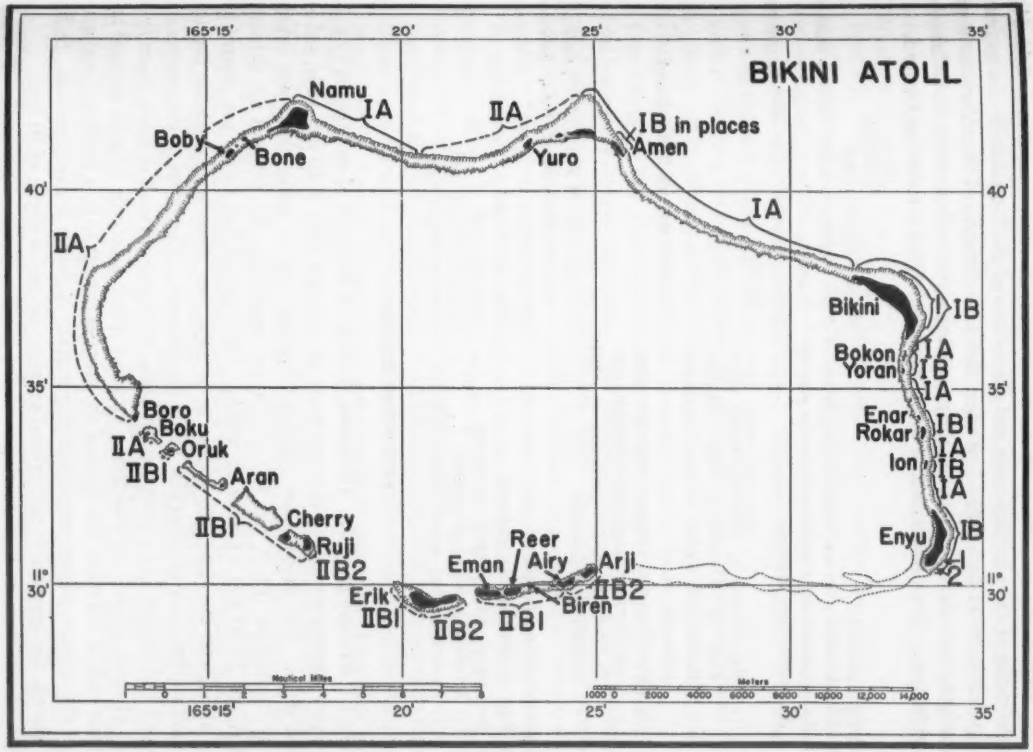



FIGURE 3—*Sketch map showing the distribution of main reef types*
Refer to Key (text). Outline from H. O. 6032.

Being in direct contact with the open sea it receives a constant supply of water rich in food and nutrient salts. It supports the greatest concentration of living organisms and to a large extent it controls the environments of the reef zones that lie behind it. The character of the marginal zone and the nature of the assemblages of organisms that live on it are in turn dependent upon the characteristics of the submarine slope upon which it stands, its relation to the prevailing waves and currents, and other ecological factors. The distribution of the various reef types is shown in Figure 3.

| | *Occurrence* |
|---|---|
| I. Strongly grooved | Wave—windward side |
| A. *Lithothamnion* ridge low, uncut by the grooves | Areas between islands concave seaward |
| B. *Lithothamnion* ridge prominent, cut by grooves that form surge channels | Areas, especially off islands, convex seaward |
| (1) Ridge formed by columnar bosses separated by intersecting surge channels, roofed over locally to form room-and-pillar structure | Northeast reef off Bikini Island |
| (2) Ridge formed of elongate buttresses separated by narrow surge channels roofed over to landward, and terminating in algal mounds and blowholes | Southeast reef off Enyu Island; southeast and south-southeast reefs, Eniwetok |
| II. Grooves weak or absent | Leeward side |
| A. Reef margin smoothly scalloped | West side Bikini |
| B. Reef margin made irregular by erosion | Southern reefs |
| (1) Re-entrants large and irregular | South-southwest reefs |
| (2) Re-entrants long, narrow, subequal in size | South-southeast reefs |

### DISTINCTIVE FEATURES

TYPE I. *Strongly grooved.* The seaward slopes of this type of reef are cut by well-developed grooves normal to the reef front 50–300 feet long, 3–10 feet wide, and 6–25 feet deep. The grooves are relatively straight but in some places are forked, usually downslope (Pl. 9, fig. 1). The spurs separating the grooves are flattened algal ridges 25–50 feet or more in width. The tops of the ridges are covered by living algae, and the sides also appear to be algal, but there is no evidence that the algae are roofing over or otherwise filling the grooves. The floors of the grooves are flattened and appear to be composed of algal limestone covered in some places by a veneer of sand, gravel, or boulders. In areas where a terrace extends seaward beyond the ends of the spurs the floors of the grooves appear to be a landward continuation of the terrace whose seaward-sloping surface is gently convex upward. So far as is known, the grooves do not extend seaward beyond a depth of 50 feet.

Reefs of this type are best developed on the windward side of the atoll. They are characterized by well-developed algal growth that appears to be adjusted to strong steady surf. The waves remove much material from the reef in small pieces which are transported down the grooves and down the terrace. Eroded parts of the reef are healed by organic growth, and the reef appears in near equilibrium. Since they are adjusted to strong surf, major storms have little additional effect.

TYPE I-A. *Lithothamnion ridge low.* The marginal ridge is a broad arch, sloping

gently seaward and rising only 6 inches to 1 foot above the main reef flat. With rare exceptions the grooves, though well developed, are limited to the seaward slope beyond low-tide level. The reef flat behind the marginal ridge generally supports a rich growth of corals. This type of reef is characteristic of reef segments between islands; these areas are typically concave to the sea. There are no islands to impede circulation of water, and on the windward side new supplies from the sea cross the reef even during ebb tide. This circulation is probably mainly responsible for the rich growth of corals.

TYPE I-B. *Lithothamnion ridge prominent.* The marginal ridge is a conspicuous feature rising 2–3½ feet above the main reef flat behind it (Pl. 1, fig. 1). The well-developed grooves of the seaward slope cut across the ridge to form surge channels that extend 50 to 100 feet from the reef edge. Between waves much of the water that is piled on the reef drains seaward through these channels.

Reefs of this type are best developed on convex arcs of reef, especially where islands are present. The islands restrict circulation of water, and such flats support very few living corals. The reef flats in such areas are subject to erosion.

TYPE I-B-(1). The *Lithothamnion* ridge in some areas is formed of isolated columnar bosses, in others of buttresses. The landward sides of such structures between surge channels are composed of calcareous algae that form a smooth pavement. These algae may grow laterally at low-tide level and eventually roof over the surge channels. This gives rise to a type of "room-and-pillar structure" along the margin of the reef that is similar to the underground workings commonly developed in mining operations. When the roofing-over process is nearly or quite complete it may be recognized on air photographs by a honeycomb pattern (Pl. 9, fig. 2).

This type of reef is well developed off Bikini Island, an area given special attention in connection with the bomb tests. It seems to require rather special conditions and probably is rare—though it may have been present at one time or another over a large part of the Bikini reef.

TYPE I-B-(2). The *Lithothamnion* ridge is formed of elongate buttresses separated by narrow surge channels. In areas where the surge channels are long and narrow—like miniature canyons—algae from the sides grow outward as a shelf at low-tide level eventually sealing the channel to form a tunnel. The landward end of such a tunnel may remain open as a blowhole long after the rest of the channel is closed, and, in such cases, a craterlike algal mound is built around the opening (Pl. 8, figs. 3, 4). Reefs of this type are developed on the south and southeast sides of Bikini but are even better developed on the south-southeast side of Eniwetok where the terminal mounds rise 2½ to 3 feet above the reef flat.

TYPE II. *Grooves weak or absent.* On the seaward slopes of reefs grooves are rare or absent, and no true *Lithothamnion* ridge is developed. Reefs of this type are best developed on the leeward sides of atolls, and in most places there is no evidence of a terrace.

TYPE II-A. *Reef margin smoothly scalloped.* The outline of the reef edge in some areas is straight or smoothly scalloped. The seaward slope ranges from very steep to vertical to depths of 30 or 40 fathoms. The marginal zone supports a fairly rich growth of corals and algae but rises only a few inches above the reef flat.

On reefs of this type growth apparently exceeds erosion; the reef is adapted to the

weak surf present on the lee side. The broad reef on the western side of Bikini belongs in this category (Pl. 5).

TYPE II-B. *Reef margin irregular.* The margins of many reefs on the south sides of atolls are very irregular in outline. Two main types of irregularities are described, one of which, at least, appears definitely to be due to erosoin. The southern coasts receive the most severe storms.

TYPE II-B-(1). *Re-entrants large and irregular.* The sharply irregular re-entrants in the reef margins in a few places are almost certainly due to erosion followed by collapse. These re-entrants are 25 to 500 feet or more wide and extend into the reef for 25 to 200 feet (Pl. 3, fig. 2; Pl. 10, figs. 1, 2). The floors of such re-entrants are terraces that probably are erosional but may represent a surface that antedates the reef. The floor at present is covered with debris—including large blocks, the irregular outlines of which match those of the reef edge. Debris of this sort occurs to depths of at least 100 feet.

Reefs with large irregular re-entrants occur on the south-southwest sides of atolls. They are adapted to the weak surf of such lee shores, but the severe storm waves from the south apparently cause major damage. Even in the re-entrants, however, large blocks torn from the reef edge are veneered with living corals and algae.

TYPE II-B-(2). *Re-entrants long, narrow, subequal.* The re-entrants of this type are 100 feet or more in length, very narrow, and apparently shallow (10–15 feet at most). They are closely spaced, separated by narrow algal ridges (Pl. 2). The re-entrants and ridges together form a sort of comb structure that rises from a floor 15–25 feet in depth.

This type of structure occurs on the south-southeast reefs. It may be related to the blowhole type described under I-B-(2), but erosion rather than growth appears to be dominant at the present time. The lines of fissurelike re-entrants form an erosion pattern that is hard to explain, and the present form may be determined by some earlier growth pattern.

## DEVELOPMENT OF THE MARGINAL ZONE

### OFFSHORE ALGAL SPURS AND GROOVES

One of the most striking features observable on many reefs, particularly when seen from the air, are the elongate algal spurs and associated "grooves" in the sea slope of the reef. The term "groove" is perhaps a misnomer because it implies that a pre-existing level surface has been chiseled at intervals to a lower level. Detailed observation, on the contrary, suggests that the spurs between the grooves are growth forms and that the pre-existing surface is now represented by the bottoms of the grooves and the terrace that in many places extends seaward beyond the grooved margin (Fig. 4). If this interpretation is correct, the position of the older surface under the present reef is now represented by the floors of the grooves and surge channels in the marginal zone, the bottoms of pools and caverns in the reef back of the marginal zone, and the floor of the reef flat in the *Heliopora* zone. In a section normal to the reef front, all these floors fall on a smooth curve, gently convex upward, that is continuous with the floor of the seaward terrace. The relationship of the marginal zone to the seaward terrace is sketched (Fig. 5).

FIGURE 1. BEACH ROCK ON ORUK ISLAND
The rock is a lithified beach sand.



FIGURE 2. BOULDER RAMPART, RUJI ISLAND
The slope of the rampart is about 15°, and its height about 8 feet above high tide level. The spar on the float is 6 feet long.



FIGURE 3. BLOWHOLE NEAR ENIWETOK ISLAND
The algal mound is 3 feet high and 18 feet in diameter.



FIGURE 4. ALGAL MOUND
An "unfinished" blowhole near Eniwetok Island showing dome structure of the algal mound.

SHORE FEATURES AND BLOWHOLES

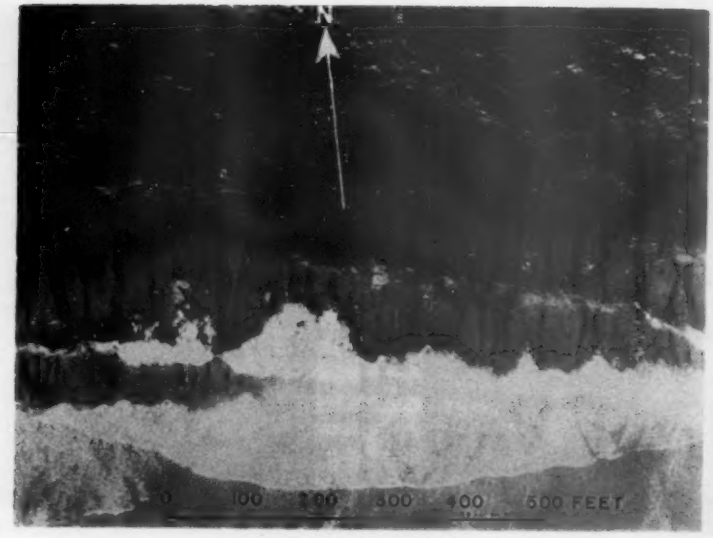



FIGURE 1. REEF TYPE I-A
Strong grooves developed on the seaward slope below low tide level. The *Lithothamnion* ridge is covered by breakers. Aerial view of northern reef of Bikini, two miles east of Namu Island. Airphoto by U. S. Navy.

FIGURE 2. REEF TYPE I-B-(1)
Irregular edge of *Lithothamnion* ridge showing buttresses or separate bosses separated by intersecting surge channels. Below them are irregular pools not yet completely roofed over. In the middle ground the honeycomb pattern outlines an area where the roofing-over is completed; this part of reef underlain by room-and-pillar structure. At the bottom of the photograph a part of the *Heliopora* zone is shown. Aerial view off the central part of Bikini Island. Airphoto by U. S. Navy.

MARGINAL GROOVES AND ROOM-AND-PILLAR STRUCTURE

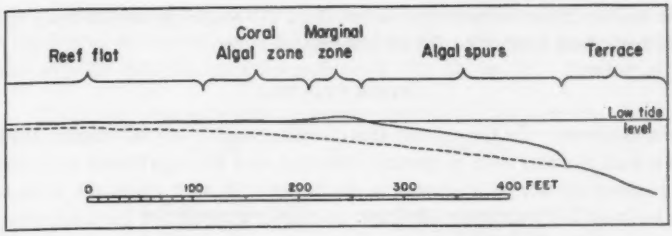



FIGURE 4.—*Cross section through marginal zone and algal spurs*

Present reef shown by solid line. Older surface, shown by dashed line, is represented by the offshore terrace, bottoms of grooves, surge channels, and deeper pools of reef flat.

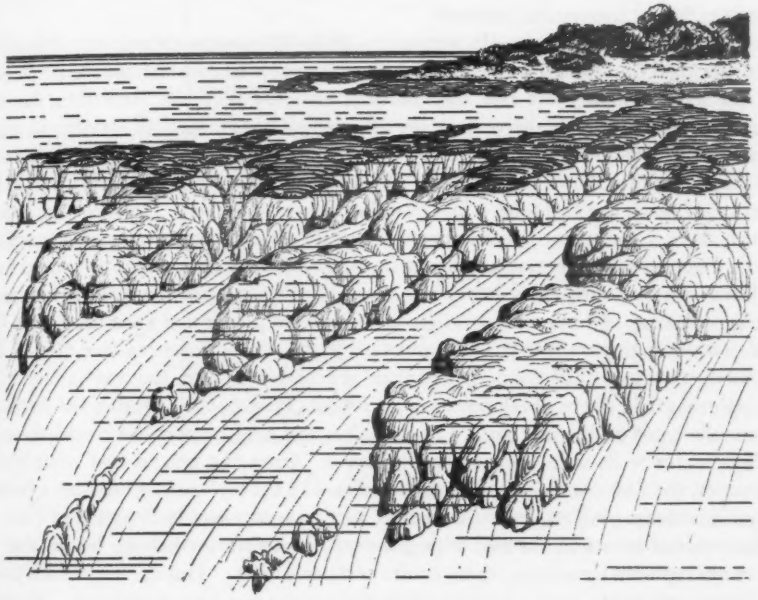


FIGURE 5.—*Sea reef off Tufa Island, Rongelap*

Sketch showing relationship of marginal zone of reef and offshore algal spurs to older surface. This illustration was sketched after swimming off a southeasterly reef that was exposed to moderate wave action. The marginal zone and spurs are not as massively developed as they would be on a reef exposed directly to the prevailing wind. Adapted by P. B. King and Margaret Austin from a field sketch by Selwyn Taylor.

Outside the marginal zone, algal growth is most vigorous on the tops of the spurs where wave motion is strongest. On the bottoms of the grooves at a lower level, water movement is dominantly outward, and all fragmental material broken from the reef is carried into these sluiceways. Apparently the veneer of debris slowly moving seaward effectively prevents algal or coral growth on the floors of the grooves.

The dimensions of the spurs and grooves seem to be adjusted to normal wave dynamics, and the reef margin is remarkably efficient in absorbing wave energy. Waves

that rise to 8 or 10 feet when only 50 feet from the reef edge are cut down to heights of 1 or 2 feet when they cross the *Lithothamnion* ridge.

### SURGE CHANNELS

*General statement.*—In the convex arcs of reefs, particularly off islands, the slope of the terrace off the reef edge is generally steeper, and the algal spurs with their associated grooves are normally shorter and steeper. In such areas the *Lithothamnion* ridge is built up to a maximum of 3½ feet above the general reef level, and the grooves continue through the ridge to become surge channels. These drain the reef flat of water piled up by the breakers, and in time of high surf the rush of water seaward through the channels may be great. Algal growth is best developed on the seaward faces of the buttresses and least on the floors of the surge channels because of the detritus that is transported seaward.

*Room-and-pillar structure.*—In some areas the *Lithothamnion* ridge is very irregular, and the algal buttresses on the seaward slope stand up as separate bosses 10–25 feet in diameter. Locally the reef may be built on these pillarlike structures, for the algae by lateral growth shelve over the intervening channels, thus forming an algal platform at low-tide level underlain by connecting caverns. Stages in the formation of the caverns are shown in Figure 6. The separate bosses (Fig. 6, A) grow vertically until they reach the surf zone, a foot or so below low-tide level. Here algal growth is much more vigorous for several reasons, probably including more intense sunlight, more vigorous movement of water past the algae, and the churning of the water into a foam with each advancing wave, with a resultant effect upon the dissolved oxygen, carbon dioxide, and minerals. As growth continues in the surf zone the bosses develop overhanging walls that shelve outward until some of them coalesce (Fig. 6, B) Growth continues until the surge channels are mere cracks on the reef surface with occasional pools where development is incomplete (Fig. 6, C). During the development the algae on the reef surface remain alive and healthy near the cracks through which water wells with each advancing wave. Once the cracks heal completely, however, the *Lithothamnion* on the surface dies, and the reef surface becomes a barren algal limestone. All stages of the process may be recognized in Figure 2 of Plate 9. The rounded bosses at the reef edge give way shoreward to shelving bosses that in turn are replaced by an area where cracks and pools exist. Finally, behind the pools a honeycomb pattern may be seen where the process has been completed. Even in this area some small openings a few feet in diameter still remain at the intersection of the roofed channels. By diving in these pools underneath the reef, one can see vaulted caverns 5–10 feet wide and 10–15 feet deep, brilliantly lighted at intervals of 50 feet or so by sunlight through small cracks and openings in the roof (reef floor) (Pl. 11, fig. 1). The structures resemble underground mine workings developed by the "room-and-pillar" type of mining. Areas underlain by caverns are generally smaller than areas occupied by pillars, although they are equal when conditions are favorable.

*Blowholes.*—In areas where the surge channels are narrow and 50 to 75 feet long, the pavement algae on the back slope of the *Lithothamnion* ridge may by lateral growth encroach upon and eventually roof over the landward 25 to 50 feet of the

channels. This process may continue until the openings are very small at the inner end of the tunnel so formed, and "blowholes" result. These holes may spout water 5–20 feet or more into the air after each wave (Pl. 11, fig. 2). Because of the con-

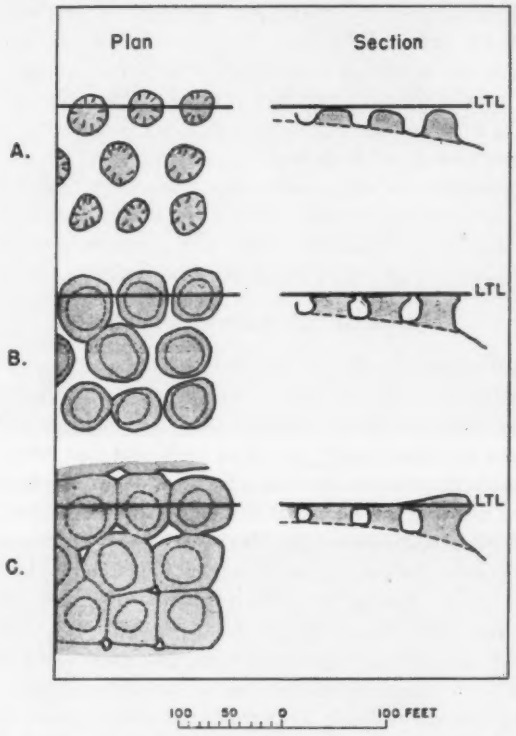



FIGURE 6.—*Diagram illustrating the development of room and pillar structure*

A. Algal bosses. B. Shelving and lateral growth in the surf zone. C. Development of reef floor underlain by room and pillar structure.

tinual supply of water, prominent live algal mounds may develop around the opening or spout. These attain a diameter of 10–30 feet and rise 1–3 feet above the reef flat (Pl. 8, fig. 3). The internal structure of one of these algal mounds not yet fully developed is shown (Pl. 8, fig. 4). On Bikini Atoll, blowholes are found off the south end of Enyu Island, but the most striking examples seen were at Eniwetok Atoll, on the reefs along Eniwetok Island. All stages of development are found there, from open surge channels at the northeast end of the island, progressing southwestward through partly roofed channels and incipient landward mounds, to completely sealed tunnels and blowholes with beautifully developed mounds. Along one segment of the reef, nearly half a mile long, every surge channel is roofed over, and the majority form active blowholes.

## EROSION OF THE MARGINAL ZONE

### DESCRIPTION

On the windward sides of atolls there are no striking features indicative of extensive erosion. The buttresses and channels of the marginal zone seem able to withstand the constant battering of the surf. The presence of debris on the floors of the surge channels and on the surface of the submarine terrace indicates, however, that much material, usually in small fragments, is broken from the reef edge, but the growth of algae is so rapid that fresh scars are rare. The development of the marginal zone suggests growth rather than erosion.

Behind the *Lithothamnion* ridge destruction takes place largely through boring organisms and solution of the reef rock. The most pitted and eroded parts are found when the flat is backed by an island. The island interferes with cross-reef circulation and thus inhibits the growth of organisms that might protect the reef surface.

### RE-ENTRANTS FORMED BY COLLAPSE

On the leeward sides of atolls the reef margin is adapted to the comparatively gentle lee surf, but it is apparently unable to withstand the attack of powerful waves generated by the major storms that come from the south. Evidences of erosion of the marginal zone are most clearly shown on reefs that face south or southwest. The waves apparently undercut the reef edge to a depth of about 30 feet. Re-entrants into the reef edge are formed at this depth, and large blocks torn from the edge lie on the floors of such re-entrants. Some of these blocks may still preserve the irregular outlines of the gap left in the reef margin by their removal (Figs. 7, 8; Pl. 10, fig. 2). In some areas, as along the southwest side of Erik Island at Bikini a major part of the reef margin has been deeply eroded in this manner. Once a re-entrant has formed, it may enlarge laterally along the marginal crest of the reef (Pl. 3, fig. 2) Though erosion seems the dominant process in such areas growth continues, and blocks separated from the reef only by narrow cracks are rejoined to the reef—temporarily at least—by a growth of algae and corals.

### COMB STRUCTURE SUGGESTING PROGRESSIVE WAVE EROSION

Along some south-southeast reefs the marginal zone is cut by numerous fissures normal to the reef edge to form a sort of "comb structure" (Pl. 2, center). The jaggedness and detailed irregularity of these structures suggest erosion, but in some places the continuations of the fissures toward the reef flat—discernible as dark markings on aerial photographs—suggest growth. Probably the erosional pattern revealed in the comb structure is determined by a pre-existing growth pattern. The fissures of the comb structure may mark the sites of former surge channels of the long and narrow variety known to occur on reefs facing southeast or south-southeast. This type of reef (II-B-2) may be derived from I-B-2 by prolonged and effective erosion.

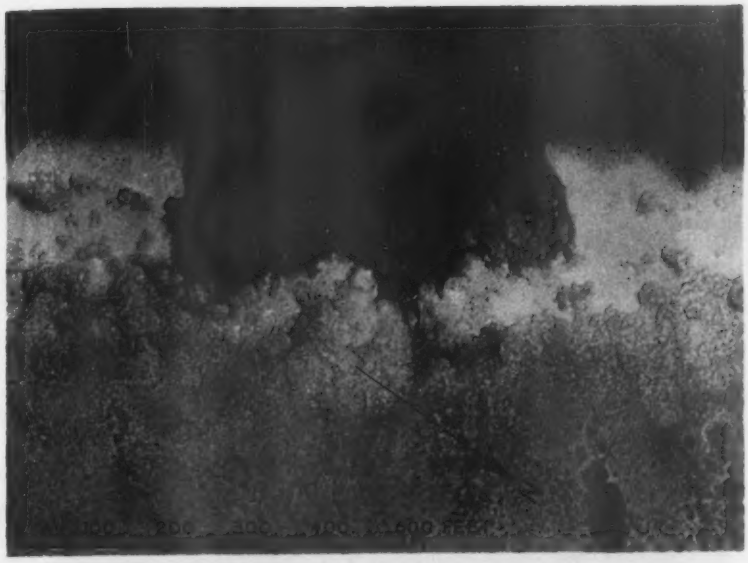


FIGURE 1. REEF TYPE II-B-(1)
Large, irregular erosional re-entrant on the ungrooved reef off the southwest side of Erik Island, on the southwest reef of Bikini Atoll. Airphoto by U. S. Navy.

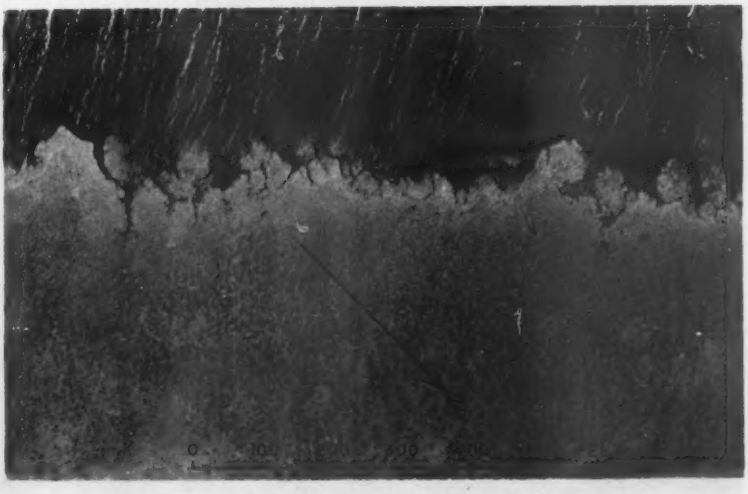


FIGURE 2. REEF TYPE II-B-(1)
Aerial view near Figure 1, showing large blocks broken from the reef edge. The original ungrooved margin has entirely disappeared in the area shown. Airphoto by U. S. Navy.

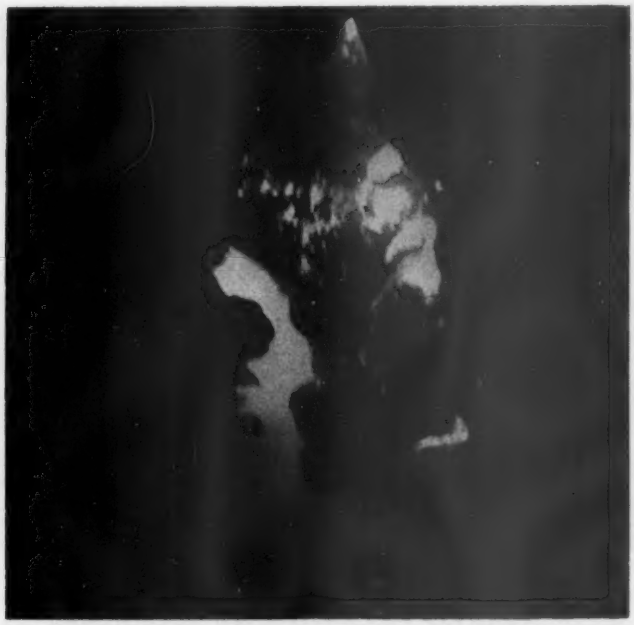


FIGURE 1. VIEW IN CAVERN NEAR REEF EDGE, BIKINI ISLAND
Photograph taken horizontally, about 5 feet below reef surface. The algal walls overhang, and join at the surface. Height of cavern about 12 feet.

FIGURE 2. BLOWHOLE SOUTHEAST OF ENYU ISLAND
In the foreground are an open pool and several crevices. The waterspout is more than 20 feet high.

REEF CAVERN AND POOLS

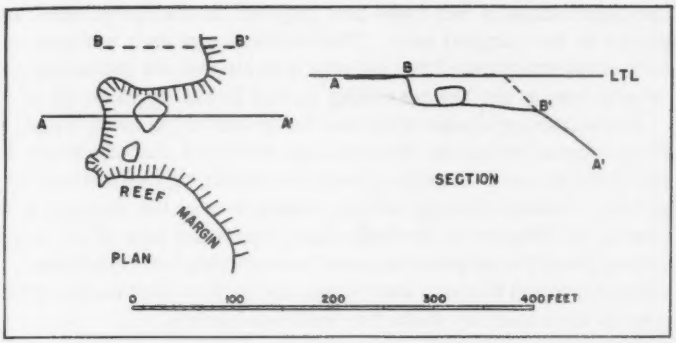



FIGURE 7.—*Erosional re-entrants on southern reefs*

AA' is a section across the reentrant in the reef and the eroded shelf or bench. BB' is a section across the adjacent eroded reef edge.

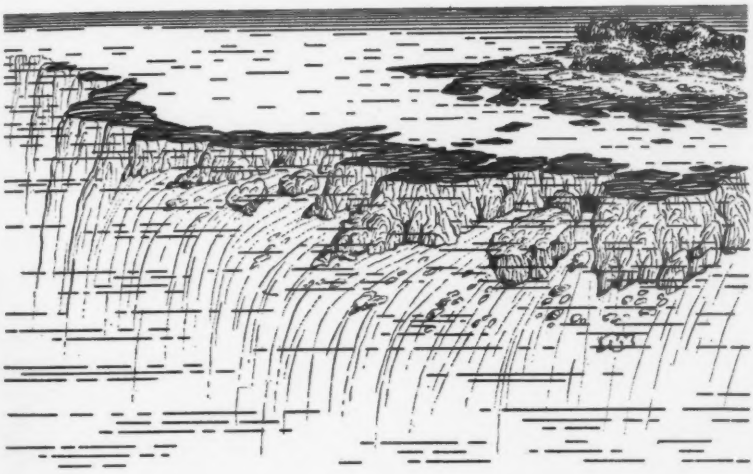


FIGURE 8.—*Lee reef off Burok Island, Rongelap*

Sketch showing large erosional reentrants in the reef edge on a southwest reef. The large blocks torn from the margin are resting on the eroded bench in about 30 feet of water. At the extreme left erosion has not occurred, and the reef margin (type II-A) dips nearly vertically to depths of more than 30 fathoms. Adapted by P. B. King and Margaret Austin from a field sketch by Selwyn Taylor.

## SUMMARY

The existing reef appears to have grown upon an older surface that extends seaward beyond the present reef margin and appears to rise landward to within a few feet of the present reef flat. The marginal zone of the reef on the seaward side is characterized by a profuse growth of algae and corals that forms peculiar structures of several kinds. In general this profuse growth does not extend deeper than 35

feet. The classification of reef types here proposed is descriptive based upon the characteristics of the marginal zone. The distribution of each reef type at Bikini and a study of the structures of the reef edge indicate that the prevailing wave conditions in any locality are the controlling factors in the development of the reef margin. In maintaining a balance between the growth of the living margin and the destructive energy of the waves, structures are developed that effectively dissipate the energy of the waves. On reefs exposed to constant high surf, strong buttresses and large surge channels develop, and the roofing over of the channels at low-tide level increases the efficiency of the baffle that absorbs the force of the waves. On lee reefs where the surf is comparatively weak, smooth margins are developed. These margins are not adapted to strong wave action, and sections that receive storm waves from the south are marked by distinctive erosional features.

U. S. Geological Survey, Washington 25, D. C.
Manuscript Received by the Secretary of the Society, September 3, 1947
Published by Permission of the Director, U. S. Geological Survey

FIGURE 1. AIRPLANE VIEW OF MOUNT MISTI (19098 ft.) AND AREQUIPA
From the Southwest. A few miles left of this view is Mount Chachani, an older and more dissected volcano. (By courtesy of Pan-American—Grace Airways, Inc.)

FIGURE 2. MOUNT CHACHANI (19931 ft.) FROM THE SOUTHEAST, AND THE VALLEY OF CHILI RIVER AT AREQUIPA.
View taken by a local photographer.

VOLCANOES NORTHEAST AND NORTHWEST OF AREQUIPA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 879-893, 5 PLS., 1 FIG. SEPTEMBER 1948

# INCANDESCENT TUFF FLOWS IN SOUTHERN PERU

BY C. N. FENNER

## CONTENTS

Page

Abstract ..... 879
Introduction ..... 880
Situation ..... 880
General characteristics of incandescent tuff deposits ..... 880
Cordillera tuffs ..... 883
Distribution of white sillar ..... 885
Distribution of salmon sillar ..... 886
Sillar ocioso ..... 887
Abnormal tuffs ..... 887
Areal extent of sillars ..... 889
References cited ..... 892

## ILLUSTRATIONS

Figure Page

1. Map of Arequipa region and adjacent territory ..... 881

Plate

1. Volcanoes northeast and northwest of Arequipa ..... Frontispiece
2. Cordillera tuffs (sillars) ..... 880
3. Relationship of tuffs and gneisses ..... 881
4. Cordillera tuffs and Arequipa sediments ..... 888
5. Features of Cordillera tuffs and lava flows ..... 889

## ABSTRACT

The present paper, the first of two closely related articles on the geology of the Arequipa region in Peru, concerns itself with the tuff deposits of the region as a volcanologic phenomenon. Their origin, mode of occurrence, and character are described. The second paper will deal with climatic conditions during the Pleistocene as evidenced by their topographic effects upon the tuffs.

The tuff deposits are the result of a series of fragmental outbursts of rhyolitic lava, similar to that which occurred in the Valley of Ten Thousand Smokes in Alaska in 1912. The period of most of the outbreaks was apparently early Pleistocene.

In the Arequipa region the deposits are of great volume, but they are only part of a much wider field that extends to the northwest and southeast. Altogether they appear to be one of the greatest known examples of this form of eruption.

Two principal types and several minor varieties are present. The older of the two is perfectly white; the younger is salmon-colored. This difference does not appear to be much more than superficial.

## INTRODUCTION

In December, 1938, and the early part of 1939, several months were spent in a reconnaissance of the country surrounding Arequipa, in southern Peru, and in working out an explanation of certain geological features. Previously this region had been little known geologically except for generalized descriptions of the most prominent features. The true nature of the enormous tuff deposits found here had not been recognized, nor their effect upon the topographic development. In this work a set of the topographic maps of the Arequipa region, based on surveys by engineers of the Peruvian Army—Servicio de Ingeniería—III Division (Comisión del Servicio Geográfico), 1929,—on a scale of 1 to 20,000, were of the greatest help. These were kindly obtained for me by Major Carlos Nicholson, at that time Professor of Geography in the University of Arequipa. For this and many other kindnesses I am greatly indebted to Professor Nicholson.

## SITUATION

Arequipa is located at latitude 16° 24′ south, longitude 71° 33′ west, according to observations at the former Harvard Astronomical Observatory in Carmen Alto, a near-by suburb. The elevation is 7550 ft. (railroad station) (Fig. 1).

The railroad running inland from Mollendo on the coast to Arequipa is barred from direct approach to the city by the Cerros de la Caldera (Kettle Hills, Fig. 1). The railroad bears toward the west, and enters the Arequipa basin through a wide gap in the Cerros, through which Chili River and Vítor River flow. Chili River is closely approached by the railroad at a point near the station Quishuarani. Here the gorge of the river is bounded on the north side by steep cliffs of salmon-colored tuffs, which have the characteristic appearance of the tuff deposits that I had previously studied in the Valley of Ten Thousand Smokes in Alaska.

From Quishuarani the railroad follows the Chili for several miles up-stream. On the south side are the steep, bare slopes of crystalline geneiss of the Cerros de la Caldera, on the north the river, and just beyond the river the cliffs of tuff, rising several hundred feet to an even rim (Pl. 2, fig. 3).

This article will describe these tuffs in detail, as representing a certain phase of volcanic activity. Their important bearing on Pleistocene and Recent physiographic development of the Arequipa region will be taken up in the succeeding article.

## GENERAL CHARACTERISTICS OF INCANDESCENT TUFF DEPOSITS

Only in recent years have tuff flows of this type been recognized and studied. Probably they are much more common than is usually realized. There seems to be still some lack of knowledge of their diagnostic features and the exact mode of their formation. Among the examples studied in various parts of the world, differences in structure and texture have been found and in some instances appear to be great, but all stages of transition between them occur, and fundamental resemblances are greater.

Contributions toward the recognition of tuffs of this origin were made many years

FIGURE 1. QUARRY IN WHITE SILLAR
On the pampas several miles northwest of Arequipa.

FIGURE 2. QUEBRADA AÑASHUAICO, IN ITS LOWER COURSE
View northerly toward Mount Chachani. Depth about 350 feet. Erosion plane at top. Walls are mostly salmon-colored tuff, but darker material on left side is a large remnant of Arequipa sediments that formerly filled the gulch.

FIGURE 3. VIEW WESTERLY DOWN CHILI RIVER
Near town of El Huaico. Salmon sillar, with planed surface, faces river, gneisses of Cerros de la Caldera at left. In center, an outjutting spur of gneiss (Cerro Alhaja) has been cut off from Cerros by superposed erosion of Chili River.

CORDILLERA TUFFS (SILLARS)

FIGURE 1. OUTLET OF QUEBRADA AÑASHUAICO
Gneisses of Cerros de la Caldera in background; in right foreground a spur of the Cerros has been cut off by Chili River.

FIGURE 2. BASEMENT BEDS AT UCHUMAYO
Dislocated sillars are underlain by pumice beds and gneissic rubble resting on steeply sloping shoulder of gneiss.

FIGURE 3. SAME AS FIGURE 2.

ago by several writers, and formed a basis for the eventual understanding of their mode of formation.

By courtesy of American Geographical Society (Johnson, George R., 1930, p. 17)

FIGURE 1.—*Map of Arequipa region and adjacent territory*

Iddings (1899) was the first to speak of "welded tuffs" and to describe examples found in Yellowstone Park, but he appears to have had little inkling of their mode of emplacement. Lacroix (1904) and Anderson and Flett (1903) were probably the first to recognize that an incandescent mass of finely fragmental material, erupted from a volcanic vent, may plunge down a mountainside as a torrent, and come to rest at its base. The terms "nueés ardentes" and "glowing clouds" were applied to these phenomena, but though these clouds are spectacular and terrifying, to call the whole group of actual manifestations "nueés ardentes" tends to place the emphasis on the relatively superficial clouds that rise from the plunging mass rather than on the mass itself, and has evidently caused some misunderstanding of what takes place. Some

of the features later found to be most characteristic of deposits of this origin, such as recrystallization and induration, were poorly developed in the deposits they described, perhaps because the initial temperature was not extremely high or because most of the material continued on into the sea. An especially important feature not prominent in the West Indies because of the conditions there, is the ability of such tuff flows to spread widely over level or gently inclined surfaces. Lacroix on the one hand and Anderson and Flett on the other held different opinions as to the mechanism of the discharge from the volcanic orifice and the importance of initial explosions and of the mountain declivity in providing the energy for the torrential descent. Anderson and Flett believed that in these avalanches "gravity did the work and supplied the energy. There was no explosion" (p. 507–512). Lacroix held that although gravity was an important factor, an initial explosion was necessary to give impetus (p. 302). Apparently neither fully recognized that the inherent properties of the mixture of solid particles and evolving gases were sufficient to impart a spreading tendency equal to or greater than that of a mobile liquid. Anderson and Flett came near to this conception in stating that "from a spongy froth the mass changed to a cloud of particles . . ., each surrounded on all sides by films of expanding gases, and thus the mixture of the ingredients from the first was perfect. Around each grain of dust there was a film of gas" (p. 507).

These observers in the West Indies marked a great advance by recognizing a previously unknown type of eruption, but these deposits were a somewhat aberrant form of incandescent tuff-flows and did not give sufficient basis for the recognition of the many deposits of this origin that occur in other places. The tuff deposits in the Katmai region in Alaska, poured out in the eruption of 1912, provided additional information, which has been found to be widely applicable.

Here the deposit covered the bottom of a great valley and came to rest with a surface of such gentle inclination that the first observers interpreted it as a mud flow (Griggs, 1918). Subsequent study showed that this idea was incompatible with the high temperature demonstrated by the charring and ignition of vegetation and with the primary and secondary characteristics of the deposit. (Fenner, 1920, 1923, 1925, 1937). It was also evident in this area that ejecta thrown high into the air as another manifestation of the same series of eruptions formed deposits of very different character. In their passage through the atmosphere they lost their heat to such a degree before reaching the ground that underlying vegetation was not charred, and the deposits were wholly incoherent. The view had been held that tuff deposits in general had originated in violently explosive ejections. Here the differences between the two types were clear.

The essence of the explanation for the deposit in the Valley of Ten Thousand Smokes was that a magma, highly charged with dissolved gases, rose to the surface through newly formed fissures in the valley floor. Instead of being thrown violently into the air, it "boiled over", and rushed down the valley in an incandescent flood of disrupted, minute particles, buoyed up by the intermingled gases, so that it followed depressions like a flood of water.

The characteristics of this tuff flow were so definite and the differences from tuff deposits of other origins so well marked as to set it off as a type—the Katmaian tuff-

flow type—and observers have been enabled to distinguish similar deposits elsewhere. In recent years a number of examples have been recognized. As a general rule, their features differ little in fundamentals from the Katmai flow, and this deposit has been taken by writers as a standard of reference. Some of the best described and most interesting examples are those studied by Marshall (1935), Gilbert (1938), and Mansfield and Ross (1935).

There has been a tendency to call all such deposits "welded tuffs", but this seems to imply softening and union by heat, and the term should preferably be restricted to cases in which this process has been dominant. For those in which induration is primarily the result of recrystallization, and for those in which the fragments have little cohesion, another term is desirable. The local term "sillar" (pronounced seelyár), commonly used in the Arequipa region, has been applied in the present paper.

Induration may be brought about in two ways. At one extreme the heat of the accumulated mass is so high after emplacement that pumice inclusions collapse and the small fragments or shards of glass of the matrix become softened and are pressed down and alined by the weight of material above them. To the naked eye the result, in extreme cases, may appear altogether like a felsitic rhyolite or even a structureless obsidian, but thin sections reveal the forms of original fragments. At the same time, a minute recrystallization of shards takes place in a texture that Iddings (1899) has termed axiolitic, and Marshall (1935), pectinate. Tuffs so consolidated are properly termed welded tuffs. No examples of this extreme type were seen in the Arequipa region.

In a second form of induration, softening is absent or minor, but recrystallization by pneumatolytic exhalations is important. The borders of each shard are minutely recrystallized, and small pores between shards are lined with somewhat larger crystals. Studies made on tuffs from the Valley of Ten Thousand Smokes and from Yellowstone Park, showed that the minute secondary crystals were tridymite and orthoclase or natrosanidine. The same minerals have been found by others in similar deposits. In tuffs of the Arequipa region these same secondary minerals occur. These tuffs contain rather numerous inclusions of white pumice, many of which originally had a fibrous structure, and recrystallization has been especially active on these pumice inclusions. In some, the glass of the pumice has been almost completely replaced by the secondary minerals. From these recrystallized tuffs, which may be very firmly indurated, one may pass to types in which the constituent fragments have been little changed, but in which there is nevertheless considerable cohesion, so that steep or vertical walls of the masses persist. A columnar structure is a common characteristic of all types.

## CORDILLERA TUFFS

The name, Cordillera tuffs, is given here to all the deposits of incandescent tuff-flow origin of this region.

In an article by Hatch (1885–1886), a collection of rocks from the Arequipa region, made by A. Stübel in 1877, has been described. The "old quartz-bearing pumice-tuff formation of Arequipa" is said (Hatch, p. 357) to extend from Arequipa to the foot of Misti. This is evidently the Cordillera tuff of the present article, of which the

actual extent is immensely greater. Stübel and Hatch go astray in conclusions regarding its origin and make-up. At the time of their work tuff-flows of Katmaian type had not been recognized, and the petrography is misinterpreted. A "porcelain-like" condition is ascribed to silicification by hot springs. By boiling a sample with a a $K_2O$ solution, 21.09 per cent $SiO_2$ was dissolved. This porcelainlike material was said to appear under the microscope as a finely fibrous, colorless substance, thought to be chalcedony. Actually it is the minutely crystalline aggregate of tridymite and orthoclase.

In the Arequipa region, as in other places where Katmaian tuffs occur, it is only by recognition of their flow origin and, consequently, their practical restriction to valleys, that one can understand their mode of occurrence, and proceed from this to an orderly interpretation of relations and sequence of events.

The Cordillera tuffs vary from pure white to reddish salmon. The white variety (*sillar blanco*) appears to be generally older. The characteristic minute recrystallization of the groundmass, and especially of pumice inclusions, by pneumatolytic action is well advanced. This has produced induration, sometimes to a high degree, but the original fragmental character is easily recognized microscopically. The salmon or reddish-brown varieties (*sillar colorado*) are occasionally recrystallized and indurated, but usually, though they are coherent enough to stand up as cliffs, they are easily broken by the hammer, and sometimes they are hardly more firm than sand. Softer varieties are known in the Arequipa region as *sillar ocioso*—useless sillar. In the Valley of Ten Thousand Smokes similar types are common.

In all varieties of these Cordillera tuffs, phenocrysts are sparse. The principal phenocryst is plagioclase, which commonly is twinned and faintly zoned. The indices correspond to oligoclase. A little biotite is present and sometimes grains of iron ore. Other dark minerals are rare. The plagioclase is often in broken fragments. At times these lack both twinning and zoning, and then closely resemble quartz, especially as the index is about the same. Some phenocrysts of this sort, however, were definitely found not to be quartz, and nothing could be positively identified as quartz. No embayed phenocrysts, or phenocrysts with conchoidal fracture or with bipyramidal terminations, such as are common with quartz in rhyolitic rocks, were seen. Hatch (1885–1886 p. 357) speaks of many fragments of quartz in these tuffs. I could not confirm this, and believe that quartz is absent or rare. Nevertheless a highly siliceous composition is demonstrated by the abundance of secondary tridymite. The secondary orthoclase has an index of about 1.525 and may be natrosanidine. The composition is that of a rhyolite or soda rhyolite. In spite of recrystallization, lunate and cuspidate forms of glass shards are preserved. Small bits of pumices and andesites are found in the thin sections.

To the inhabitants of Arequipa the tuffs have been very useful as a building material. They are so cheap and easily worked, and so durable in this frost-free climate, that they have been used in even the poorest structures for several hundred years. The best variety is almost snow-white, and has been so generally used that Arequipa has become known as La ciudad blanca, the white city. This rock, though porous and composed originally of minute, incoherent shards of glass, has been so recrystallized and indurated that when a block of it is struck with a hammer it gives out a clear,

bell-like note. Other varieties, generally those of a salmon color, are less firmly consolidated.

The tuff flows were of great volume. There were doubtless a considerable number of them, but only rarely is there definite evidence of separate flows. Faint horizontal lines such as are visible in figure 1 of Plate 2 are probably indicative of more than one flow, and in the great canyon of Chili River north of Arequipa (Pl. 5, fig. 2), the evidence of a succession of flows is more definite. In deep quebradas (gulches) in a region several miles southwest of Arequipa, evidence of another kind is found. In their upper (northeastern) courses the walls of these quebradas are white sillar; in the lower courses, salmon sillar (sillar colorado). The contract is concealed by talus, but white sillar is believed to be the older. Along these quebradas a remarkably developed peneplain is exposed, cutting across both white and salmon sillars, and upon this surface *sillar ocioso* has been deposited. The amount of erosion between flows of the different types indicates hiatuses of probably hundreds or possibly thousands of years.

## DISTRIBUTION OF WHITE SILLAR

A few miles north of Arequipa, on the lower slopes of Mount Misti that face westward toward Chili River, there are quarries of white siller (Pl. 4, fig. 1), now abandoned, which, from their extent and the amount of refuse present, were probably the principal sources of building stone for many decades. It is of fair quality, but is not equal to that in newer quarries out on the pampas several miles northwest of Arequipa, where the principal quarrying is now carried on (Pl. 2, fig. 1).

The situation of the quarries at the first mentioned site with reference to Mount Misti may be seen from figure 1 of Plate 1. Though these workings are actually very extenisve, they appear as only a small white patch near the left border of the photograph. All these lower slopes are underlain by white sillar for a long distance up the river from here. Several large, dry quebradas (gulches) extend back into the slopes of El Misti and give additional exposures. At one time all this region must have been covered with sillar to a great depth. Large remnants are found to a height of about 8400 feet on the adjacent southwestward-facing slopes of El Misti, and a probable remnant was seen from a short distance at about 9300 ft. On the basis of these figures, and judging from the low angle of repose that the similar flows in the Valley of Ten Thousand Smokes assumed (1° 08′, or a slope of 1200 ft. in 11½ miles) they probably covered the whole Arequipa basin to such a depth that they spilled over low places in the Cerros de la Caldera (Kettle Hills) on the southern border of the basin. Undoubtedly, large volumes escaped through the southwestern gap where the Chili and Vítor Rivers cut through the range.

Other extensive exposures of white sillar appear along several large quebradas west and southwest of Arequipa, especially along the upper parts of Quebradas Río Seco, del Crucero, Escalerilla, Añashuaico, and del Huaico.

At the eastern border of the basin large remnants of similar white sillar are present near Chihuata, 10 miles from Arequipa, at an elevation of about 9000 ft. Near the Rescate Mine, 14 miles to the southeast, this sillar occurs up to 8300 ft.

## DISTRIBUTION OF SALMON SILLAR

After the deposition of white sillar, a great part of it was removed, during a period of erosion. The thickness of what is left is unknown, for sections of 400 ft or more in quebradas do not expose the underlying surface.

The western part of the basin must have been deeply covered with white sillar, as shown by the exposures just mentioned along the upper parts of various quebradas, but in their lower courses they pass from white sillar to salmon sillar (Pl. 2, fig. 2). The contact is concealed by talus, but there is a sudden change from one variety to the other. There is no corresponding break in the topography of the rims—evidently a planation had occurred after both had been emplaced.

Some of this salmon sillar shows evidences of pneumatolytic activity, to a degree unusual for this variety. Pumice inclusions have been recrystallized to aggregates of secondary minerals, and they contain a small quantity of a black, metallic mineral, probably fumarolic iron ore. Salmon sillar continues southwestward down the quebradas mentioned to their outlet at Chili River (Pl. 3, fig. 1). Along this part of the Chili the bluffs are composed of salmon sillar (Pl. 2, fig. 3).

Where Quebrada Añashuaico opens out into the valley of the Chili (Pl. 3, fig. 1) the thickness of sillar exposed in the bluffs near by (not shown in the photo) is nearly 400 ft. In this vicinity the sillar and associated beds rest upon gneiss (Pl. 3, figs. 2, 3). From the relations shown, the course of events seems clear. The tuff flow was immediately preceded by a heavy fall of pumice. As the nearest volcano from which this could have come (Mount Chachani) is 18 miles to the north, great violence of eruption is indicated. After the pumice, the tuff flow followed in great volume. The steepness of the hillside upon which the ejecta came to rest caused slumping and dislocation (one or more times), which involved the rubble, the pumice, and the sillar. Although the slumping presumably occurred almost immediately after emplacement of the tuff, pneumatolytic emanations given off by it had already transformed it from a collection of loose particles to a firmly coherent mass that broke into blocks.

In northwestern areas of the Arequipa basin, in the general region of Yura and the Chocolate quarry (Pl. 5, fig. 1), the pampas show beds of bowlders and gravel overlying an irregular surface of salmon sillar.

A little to the north of Arequipa, both white and salmon sillar appear along the valley of Chili River. Salmon sillar is later (Pl. 4, figs. 1, 2). Its surface is hummocky, and in most exposures it is covered with a thick series of sedimentary strata (Arequipa beds), composed of bowlders, gravel, sand, and pumice. At Carmen Alto, the former astronomical observatory of Harvard University is situated on a high knoll of this sillar at the edge of the valley.

The Arequipa sediments mentioned above will be more fully described and interpreted in the succeeding article. They are believed to be outwash deposits of morainal detritus, which, during the glacial period, filled quebradas and river channels that had been incised in the sillar. At a later period most of them were carried away.

Several miles north of Arequipa, Chili River emerges from a profound canyon with precipitous walls. Sections of salmon sillar exposed here were estimated to be 500 ft. thick (Pl. 5, fig. 2). They reach an elevation of at least 9000 ft. These flows,

like those of the earlier white sillar, must originally have covered the Arequipa basin deeply, but the greater part has been removed.

All these occurrences of both white and salmon sillar contain scattered inclusions of white pumice and angular pieces of andesitic lava of various sorts, ordinarily 3 or 4 inches in diameter. Many observations were made to see whether variation in size of inclusions would give a clue to the source from which the sillars had been erupted, but no consistent change was noted. Mounts Chachani and El Misti seem probable sources, but Pichu Pichu, Ampatu, and various volcanoes lying to the rear (northeast) of the front line of volcanoes seem possible sources. The sillars have a much greater extent than is included in the area of the Arequipa basin, and must have been ejected from a number of vents.

In the thick deposits in Chili canyon, intercalations of pumice beds occur, and the overlying sillar for a few feet above such strata includes large quantities of strewn pumice, as if the flow of sillar, passing over a bed of pumice, had swept up some of it and incorporated it in its lower mass, but the force of the disturbance was not sufficient to remove the whole bed. Much of the pumice in these strata is in irregular rather than rounded bumps, implying not much transportation or weathering.

## SILLAR OCIOSO

The sillar to which this term is commonly applied in the Arequipa region resembles salmon sillar (sillar colorado) in color but is incoherent. Probably typical salmon sillar that was originally firmly consolidated may, by weathering, lose coherence, but there are certain occurrences of sillar in which a firm condition seems never to have been attained. These were laid down on the previously mentioned planed surface that truncates typical white and salmon sillars. From this relation to the earlier flows, it appears that after the great planation, there were renewed flows of sillar, salmon in color but pulverulent. These apparently did not amount to the tremendous volumes of the earlier white and salmon outbursts.

## ABNORMAL TUFFS

The three types of sillar described, or small variations of them, are of great magnitude in the Arequipa region. Their characteristics are so definite and correspond so well to incandescent tuff flows in other regions that no uncertainty was felt in interpreting their origin. As will be explained more fully in the later article, their subjacent position in relation to certain important outwash deposits believed to be of morainal derivation (Arequipa sediments) makes it appear that they are of pre-glacial age, probably early Pleistocene. There are, however, other tuff deposits, much smaller than those described, that differ from them in greater or less degree and are believed to be of intra-glacial age. They depart from the well-characterized types chiefly by their inclusion of large amounts of extraneous matter, such as rounded or angular andesitic bowlders or gravel. Many have a dull or earthy look rather than the fresh appearance of the normal tuffs, but they vary considerably. Some suggest morainal material in which a large amount of ground-up tuff or pumice has been incorporated (and certain true morainal deposits appear to be so constituted).

Others resemble ordinary salmon sillar. The similarity to moraines on one hand and to sillar on the other presented a puzzling problem, on which light was eventually thrown by the recognition of certain significant characteristics.

Wherever definite lumps of fragile pumice or sillar were found in any of them a morainal origin was regarded as excluded, but the problem of accounting for their differences from normal sillar remained.

In several occurrences, field relations showed that they were deposited during the period of the Arequipa intra-glacial sedimentation rather than in pre-glacial time. In normal pre-glacial sillar, whether hard or soft and whether white or salmon-colored, the associated lumps of pumice are perfectly white and contain only a very few minute crystals of hornblende or biotite. This is true of the sparse lumps of pumice that are distributed through the normal sillars everywhere and of the basal beds of pumice at Uchumayo (Pl. 3, figs. 2, 3) and the intercalated layers among the sillars in the upper canyon of the Chili (Pl. 5, fig. 2). In composition they are highly siliceous, rhyolitic pumices.

In the abnormal tuffs, included pumices are sometimes white but more often are dingy gray or buff or banded, and carry more of the mafic minerals. They are of andesitic or near-andesitic composition. In this respect they fall in line with late pumices that occur apart from sillars, intercalated in Arequipa sediments or in beds of such recent deposition that they lie on the very surface of the ground. There is a definite tendency for the composition of pumices to vary with age from rhyolitic in all early pumices to andesitic in most of the later ones. This is so general that a presumption seems warranted that any tuff in which the associated pumices are rather mafic is not pre-glacial but intra-glacial or post-glacial

These criteria seem applicable to most of the abnormal tuffs. Though not wholly rigorous in their implications, they point quite plainly to the manner of origin. Certain deposits looked abnormal in still other respects, but sufficient time was not found for adequate study of these. Deposits near the mouth of the great Chili canyon north of Arequipa come within this category.

The abnormal tuffs for which an explanation has been deduced occur at intervals in an arc at the southwestern foot of El Misti for a distance of about 7 miles, from Quebrada Nacaco at the north, southeastward to well beyond Quebrada Bateones. They are thought to have been mud-flows, generated by outbursts of incandescent sillar from a summit crater (probably from El Misti), that plunged down mountain slopes covered with glaciers and moraines, melted much ice, and gathered up foreign material.

Good examples of these heterogeneous deposits occur in Quebrada Nacaco, one of the large quebradas that descend the southwestern slopes of Mount Misti north of Arequipa. Near the mouth of this quebrada the lower walls are formed of ordinary salmon sillar, whose irregular surface is covered in normal manner with stratified gravel and bowlders—the Arequipa sediments. Resting upon them is a large heterogeneous mass of different aspect. It is an unsorted jumble of andesitic bowlders and pebbles, many of them angular, with lumps of pumice, in a sandy matrix apparently composed largely of crushed pumice and sillar.

Farther up this quebrada, a mass of heterogeneous material cuts out a thick series

FIGURE 1. LOOKING SOUTHERLY ALONG CHILI RIVER
In middle distance, old quarries in white sillar. Landscape shows development of terraces and pediments on Cordillera tuffs and Arequipa sediments at southwestern base of Mount Misti.

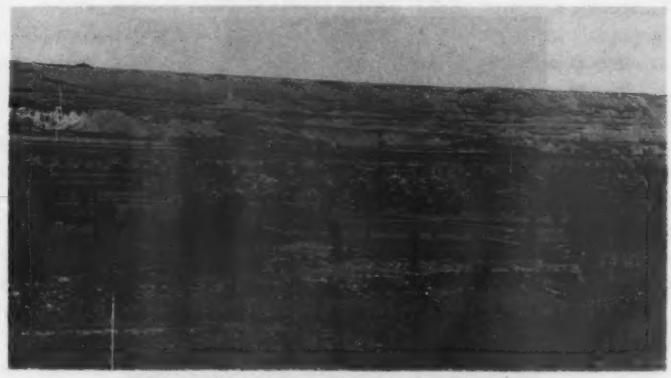


FIGURE 2. ON WESTERN SIDE OF CHILI VALLEY
Wall of Arequipa sediments, about 200 feet high, overlying eroded surface of sillar colorado. (Strongly marked bands are artificial terraces.)

FIGURE 3. IN QUEBRADA NACAGO
Arequipa sediments cut out and replaced by mass of unsorted material (upper left).

FIGURE 1. EROSION FORMS IN SILLAR COLORADO
Northwestern part of Arequipa Basin. At left and in lower right corner are remnants of a large flow of andesite from Mount Chachani.

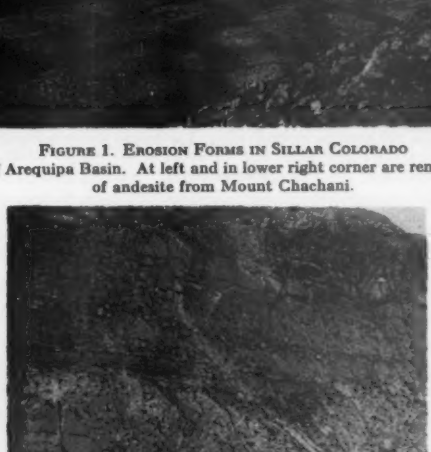


FIGURE 2. GREAT CANYON OF RÍO CHILI
North of Arequipa. Walls of sillar, estimated to be 500 ft. high, capped by Chili andesite, a lava which followed an old, high channel of the river. Banded structure of sillar indicates several great flows.

of Arequipa sediments (Pl. 4, fig. 3), but in turn is overlapped by bedded strata (not shown), lying farther to the left. It appears that here, in the midst of the deposition of the Arequipa sediments (glacial outwash deposits), erosion removed a large section, and the gap was filled with heterogeneous material, after which orderly sedimentation was resumed. In the lower courses of adjacent quebradas cutting these southwestern slopes of El Misti, deposits of unsorted detritus of quite similar appearance occur.

In Plate 1 figure 1 there is apparent a strong suggestion that the large quebrada in the center (Quebrada de San Lázaro) has been the source of a great apron of detritus, which spread outward from the mouth. This apron is of somewhat complex makeup, but most of it is believed referable to one or several mud-flows of the origin just described.

The portion of the apron at the left (northwest) of center in figure 1 of Plate 1 is the Pampa Polanco. The surface material of this resembles brown sillar. It contains much pumice, some of which is almost black. Many bomblike spherical lumps, 3 to 6 inches in diameter, are completely sheathed in a dense crust (largely glassy), enclosing a frothy interior. In this area, occasional exposures in quebradas show that beneath the brown deposit lies typical white sillar, irregularly eroded.

This surface material of Pampa Polanco contains only a small amount of dense lavas and a rare bowlder, 2 or 3 feet in diameter, but in the deep quebrada shown at the far left of Plate 1, figure 1, walls of the same tuffaceous material, 50 or 60 feet high, contain many large rounded bowlders. A faint but definite banding divides the deposit here into several thick layers. At one place this heterogeneous material is underlain by a 12-inch layer of ragged pumice, beneath which is a bed of bowlders whose base is concealed; at another place the underlying material is a succession of silty layers, abruptly cut out at one end.

These relations show the complexity of the history here, and the meaning of only the major events can be suggested. The great deposit of white sillar that, much earlier, had covered this area, had been eroded to an irregular surface. Arequipa sediments of intra-glacial age were deposited, and partly washed out. Minor events occurred, and then one or several great mud-flows, originating in the manner suggested above, came down Quebrada de San Lázaro and spread widely over this area. Without going into further details, there is evidence that in the later history of this apron torrential streams coming down from El Misti along the Quebrada de San Lázaro and adjacent quebradas cut deep trenches in the mud-flow deposits. Then Arequipa sedimentation was resumed and the trenches were filled, and finally most of the sediments were removed. Pleistocene history in the Arequipa region had many vicissitudes, climatic and other. Further examples will be given in the succeeding article.

## AREAL EXTENT OF SILLARS

When my studies of the sillars of the Arequipa Basin were begun, these tuffs were supposed to be of limited extent and to have come from a local source and an endeavor was made to find their limits or to find in their increased thickness in some direction

or in the larger size of inclusions of dense rocks or pumice, the point from which they had spread. As the work progressed the area covered by sillars became wider and wider, without any positive indication of source. Finally, a new conception of their areal extent was gained when I left Arequipa by plane for Lima, 475 miles to the northwest. Immediately after leaving the airport (situated on an outwash gravel plain near the base of Mount Chachani, a few miles northwest of Arequipa) the course lay over the sillar country shown in Plate 5, figure 1, and this characteristic landscape continued for a long distance, becoming more and more deeply dissected. The height of the airplane above the general surface was judged to be not more than 1000 feet, and in the clear atmosphere details of the quebradas and their sculptured sides were very sharp. Everywhere the irregular pampa surface was intersected by trenches. No solid rock appeared, and the great depth of quebradas gave the landscape a fantastic appearance. The surface seemed to be absolutely bare of vegetation except for a little occasionally at the bottom of quebradas. The sillar was generally salmon-colored, but broad bands of somewhat varying shades along sides of quebradas indicated a succession of flows. The depth of the quebradas gradually increased until finally they appeared to be, at a conservative estimate, 1000 to 1500 feet deep.

About an hour after leaving Arequipa airport the plane passed over a deep valley that separated the sillar region to the southeast from hills of obviously different character to the northwest, and no more sillar was seen.

Calculating from the time taken to fly from Arequipa to Lima (a few minutes under 3 hours), the time during which we were passing over what seemed to be sillar (1 hour), and the distance of 475 miles between the two cities, we find 158 miles for the distance across the sillar region. The deep valley that marks the boundary may well be at the headwaters of Ocoña River.

In a valuable pamphlet by Marsters (1912) on the Physiography of the Peruvian Andes which seems to be less well known that it deserves, numerous references are made to "mud-flows". Though small mud-flows are present in the Arequipa region and may occur also in the regions described by Marsters, there is reason to regard most of his mud-flows as Cordillera tuffs. This was my impression in traveling by airplane over part of his region, and his own descriptions seem to be in accord. At the time that Marsters wrote, the idea of incandescent tuff flows had not attained general recognition, and the widespread sheets of indurated material that they form were generally attributed to mud-flows.

A photographic view of the "Chili Valley near Arequipa", (Marsters, Pl. XXIX) shows a familiar scene. It is described by Marsters as "near the contact of the mud-flow with the edge of the foot-hills of the west slope." This is certainly the Cordillera tuff. Another photograph having the caption, "Chili-Vítor Canon and the foothills of the west range of the Andes" (Marsters, Pl. XXVIII) is easily identified as having been taken near the railroad station of Quishuarani, mentioned in this article as the place where Katmaian tuffs were first recognized in approaching Arequipa.

The common situation of Marsters' "mud-flows" along the valleys of the coastal region running down to the sea is precisely that which the Cordillera tuffs of the Arequipa region would have taken in their downward progress.

Marsters appears to have been a careful observer. He evidently was well acquainted with the region along the Ocoña River, and his descriptions are in accord with my inferences, if we may be allowed to interpret his "light-colored lavas having the form of mud-flows" as our Cordillera tuffs. Extracts from his article read as follows (Marsters, 1912, p. 240 ff.):

"The town of Ocoña is situated on the coast some seventy miles to the northwest of Mollendo. The point known as Cora Puno is located near the western edge of the West Range. It is one of the highest collection of volcanic peaks on the west of the Cordillera. Starting at Ocoña, we shall run our section across the immense plain or desert of Cuno cuno to the valley of the Chorunga, thence up the Andaray cuesta to Cora Puno and the West Range. . . . Passing inland . . . towards the north . . . we soon realize the fact that the surface of the Cuno cuno plain has once been covered in part at least by a sheet of mud-like lava . . . This light-colored sheet now persists until we reach the steep slope that passes into the Chorunga Valley. Here the Chorunga as well as the Rio Grande and the Ocoña have succeeded in cutting through, not only the soft lava referred to above and the underlying sediments, but also have deeply incised themselves in the underlying pre-Tertiary formations of the partly buried foot-hills. . . . The Chorunga simply cleaned out the mass of sediments and sheet of lava which had been dumped into it.

"Ascending the Andaray cuesta, we go but a short distance up the slope before we again see remnants of the familiar lava [tuff] sheet of the Cuno cuno plain. . . . From Andaray [which is 110 miles northwest of Arequipa] we pass up another steep slope or series of slopes, composed in part of pre-Tertiary rocks, only to find that we are again on a somewhat dissected plain which extends away to the east in the direction of Chuquibamba and north in the direction of the picturesque Cora Puno. Lithologically, this is the same type of rock [Cordillera tuff] as that which we have already seen in the Chorunga Valley, the Andaray cuesta and the inner edge of the coastal plain. It is my belief that the plains at the base of Cora Puno group of domes are but parts of the same lava plain that we have already recognized in a portion of the Cuno cuno desert."

Another section, from Ocoña to Caraveli, showing similar phenomena, described by Marsters. Here

"The pre-Tertiary foothills were only partly buried in sediments, but later they became almost obliterated by the accumulation of mud flows of great thickness and areal extent" (p. 243).

He refers again to the Ocoña-Cora Puno section and states

"Inland . . . The Tertiary sediments give way to the development of an enormous mud flow, which, on the inner edge, or escarpment overlooking the valleys of the Chorunga and Ocoña, is not far from 1,000 feet in thickness. . . . From the northeast slope of Cora Puno we have before us a gently ascending plain, with cerros appearing here and there above the general sky line. . . . The plains between the peaks are, in part, at least, made up of a light-colored mud flow. . . . Occasional small knobs of limestone protrude through this sheet" (p. 252).

Marsters finds the "mud-flows" first appearing in the northwest coastal region in the canyon of the Caraveli and in tributaries of the Atico. From these, southeastward to the Moquegua River (about 100 kilometers southeast of Mollendo, *see* his map), each of the several valleys shows these deposits. The distance along the Peruvian coast is about 275 kilometers (170 miles), from Rio Atico to Rio Moquegua.

These data give a good idea of the volume and of the area covered by his "mud-flows," which are here interpreted as our Cordillera tuffs. They were apparently derived not from one but from several volcanoes of the Western Cordillera. Petrologists who may wish to speculate on the fundamental causes of volcanism have the problem of finding any reasonable explanation as to why a number of widely spaced vents, at a certain stage in their history, poured out these highly differentiated products in a highly specialized form of eruption.

Marsters does not speak of "mud-flows" farther to the northwest than the Atico Valley, but Broggi believes that he has seen our Cordillera tuffs far to the north (Broggi, in Fenner, 1940, p. VIII). He has written as follows (translated):

"... the rhyolitic tuffs of the Arequipa type do not constitute a local case in Peru. In Cajamarca [not far south of the Ecuadorean boundary] ...; in Junín [a few miles south of Cerro de Pasco] in the famous Rock Forest, and Casapalca [not located], we have found the same rocks, covering all the others existing, and with evidence of having experienced only a very recent fluvioglacial erosion."

Rich (1942, p. 211), in an account of an aerial reconnaissance of South America, describes the aspects of the region northwestward from Arequipa. Rich's airplane course was apparently a little closer to the sea than mine. The account reads:

"... for about 140 miles we flew ... over the sloping surface of a series of pampas, ... separated by canyons 1000 to about 4000 feet deep cut by the Río Majes, the Río Ocoña, the Quebrada de Atico, the Río de Chaparra, the Río de Lomas, and smaller streams.... The steep walls of the canyons reveal perfect cross sections of a few hundred to a thousand feet or more of relatively even-bedded unconsolidated Tertiary sediments resting unconformably on an uneven surface of older crystalline and metamorphic rocks."

Rich's description, like that of Marsters, confirms my impression of the topography of the country, but Marsters' "mud-flows" and Rich's "Tertiary sediments" are believed to be the Cordillera tuffs.

If we may thus interpret them, then the observations of Marsters, of Rich, and of Broggi supplement and confirm the importance of the Cordillera tuffs as a phase of eruption among the Andean volcanoes. Previously, their character seems to have been misinterpreted and the important part they played has been far too little recognized, both the distinctive character of the eruptive manifestation that they represent and the enormous bulk of material so extruded.

The dense lavas of the Arequipa region are predominantly andesites, but the tuffs (sillars) are very siliceous rocks—rhyolites and soda rhyolites. The sillars have yielded easily to erosion; nevertheless the mass left amounts to a large fraction of that of the dense lavas. The common idea of the overwhelming excess of andesites among Andean eruptives seems to require modification.

## REFERENCES CITED


Anderson, Tempest, and Flett, J. S. (1903) *Report on the eruption of the Soufrière in St. Vincent in 1902*, Royal Soc. London, Tr., ser. A, vol. 22, pt. 1.

Fenner, C. N. (1920) *The Katmai region, Alaska and the great eruption of 1912*, Jour. Geol., vol. 28, p. 569–606.

——— (1923) *The origin and mode of emplacement of the great tuff deposit of the Valley of Ten Thousand Smokes*, Nat. Geog. Soc., Contrib. Tech. Papers, No. 1, 74 pages.

——— (1925) *Earth movements accompanying the Katmai eruption*, Jour. Geol., vol. 33, Nos. 2 and 3, p. 116.

——— (1937) *Tuffs and other volcanic deposits of Katmai and Yellowstone Park*, Am. Geophys. Union, Tr. 18th Ann. Meeting, p. 236–239.

——— (1940) *Geología de los alrededores de Arequipa*, a paper published in Arequipa for private circulation.

Gilbert, Charles M. (1938) *Welded tuff in eastern California*, Geol. Soc. Am., Bull., vol. 49, p. 1829.

Griggs, Robert F. (1918) *Ohio Jour. Sci.*, vol. 19, No. 2, p. 117.

——— (1922) *The Valley of Ten Thousand Smokes*, Nat. Geog. Soc., p. 277 ff.

Hatch, F. H. (1885–1886) *Ueber die Gesteine der Vulkangruppe von Arequipa*, Tscherm, Min. Petr. Mitt., vol. 7, p. 308–360.

Iddings, J. P. (1899) *Geology of the Yellowstone National Park*, U. S. Geol. Survey, Mon. 32, pt. II.

Johnson, George R. (1930) *Peru from the air*, Am. Geog. Soc., 159 pages.

Lacroix, A. (1904) *La Montagne Pelée et ses éruptions*, Paris.

Mansfield, G. R. and Ross, C. S. (1935) *Welded rhyolitic tuffs in southeastern Idaho*, Am. Geophys. Union, Tr., 16th Ann. Meeting.

Marshall, P. (1935) *Acid rocks of the Taupo-Rotorua volcanic district*, Royal Soc. New Zealand, Tr., vol. 64, 44 pages.

Marsters, V. F. (1912) *The physiography of the Peruvian Andes with notes on early mining in Peru*, N. Y. Acad. Sci., Ann., vol. 22, pp. 225–258.

Rich, J. L. (1942) *The face of South America; An aerial traverse*, Am. Geog. Soc., 299 pages.

64 Broad St., Clifton, N. J.

Manuscript Received by the Secretary of the Society, July 12, 1946.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 895-917, 5 PLS. SEPTEMBER 1948

# PLEISTOCENE CLIMATE AND TOPOGRAPHY OF THE AREQUIPA REGION, PERU

BY C. N. FENNER

## CONTENTS

| | Page |
|---|---|
| Abstract | 895 |
| Present Topography of the Arequipa Basin | 896 |
| Cordillera tuffs | 897 |
| The Quebradas and Arequipa sediments | 897 |
| Preliminary outline of sequence of events | 899 |
| Relations of sillars and Arequipa sediments along the Chili valley north of Arequipa | 903 |
| Unsorted detrital deposits | 906 |
| Possible methods of accumulation of unsorted debris | 911 |
| Observations on glacial geology of Andean regions by others | 913 |
| Further inferences regarding Pleistocene climate in the Arequipa region | 914 |
| References cited | 918 |

## ILLUSTRATIONS

| Plate | Page |
|---|---|
| 1. Quebradas. River terraces | 898 |
| 2. Arequipa sediments | 899 |
| 3. Canyons cut in sillars, in andesite, and in morainal material | 904 |
| 4. Glacial deposits | 905 |
| 5. Bowldery accumulations. Diatomaceous silts | 908 |

## ABSTRACT

In the reconnaissance studies of the Arequipa region, two principal subjects were investigated. One was the enormous deposits of tuff, originating as incandescent flows of Katmaian type, which cover most of the region; and the second was the climatic conditions that prevailed during Pleistocene time, and their effect upon topography. Since a large part of the evidence on climate is found in the tuff deposits the two subjects are closely inter-related, and cannot be wholly separated. Nevertheless it has been thought best to describe the tuffs in one article, immediately preceding this, as representing a certain phase of volcanic activity; and the topographic features and their significance in another. A preliminary reading of the first paper will help in the better understanding of the present paper.

Considerably before the establishment of glaciers on the high mountains north of Arequipa the lower country had been covered by a series of incandescent tuff flows. In these, deep quebradas (gulches) had been cut and a new valley of Chili River had been established before the glaciers became active. This erosion was accomplished by torrential streams, and a pluvial period preceding glaciation is indicated. Next, the gulches were filled with bowldery sediments, which also spread over the intervening pampa surfaces, and finally the sediments were almost completely removed. These later events are regarded as the consequences of the over-

loading of glacial streams with detritus, followed by still greater flow of glacial meltwaters.

The belief that during the Pleistocene there were great glaciers at work on the high mountains bordering the basin is based chiefly on the presence of huge heaps of bowlders at the bases and on the slopes of the mountains. As the importance of glaciers in this and other Andean regions has been questioned by some observers, the deposits are described in detail and the possibility of their having been formed by other processes is examined. It is concluded that no other process is satisfactory as the principal agency. Other processes doubtless co-operated, but it is believed that glaciers were the chief means by which the materials were gathered up, worked over, and deposited where now found.

## PRESENT TOPOGRAPHY OF THE AREQUIPA BASIN

Arequipa is situated in latitute 16° 24′ south, 71° 33′ west, at an elevation of 7550 feet (railroad station), at a distance of 55 miles northeast from the Pacific coast (Fenner, 1948, fig. 1). It lies in a broad valley or basin, bounded on the southwest by the Cerros de la Caldera (Kettle Hills) and on the northeast by the great volcanic mountain masses of Chachani, El Misti, and Pichu Pichu (respectively 19931, 19098, and 18583 ft. high, according to latest data supplied by the American Geographical Society). These mountains and Chili River have been of great importance in the late climatic and topographic history.

There is a fair rainfall during January, February and March in the part of the Arequipa Basin lying close to the mountains and extending over Arequipa itself, but during most of the year Arequipa is almost without rainfall, and away from Arequipa southwestward toward the ocean, even the small rainfall dimishes to almost nothing in a remarkably short distance. After a rainy afternoon in Arequipa, and while the sky is still overcast, one may look toward the southwest and see the sun sink below the cloud bank. Thus the brilliant sunsets for which Arequipa is noted are produced.

The three volcanoes mentioned, and others to the northwest and southeast of them, are distributed in an approximately straight line along the southwestern edge of the main Western Cordillera. Chili River and its upper tributaries rise outside of the Arequipa Basin, within the main Cordillera region. The river comes down between Mount Chachani and El Misti through a very narrow, deep canyon, turns slightly toward the left, and then flows almost due south across the basin until it runs head on into the Cerros de la Caldera. There it turns abruptly to the right, and follows a sinuous course along the very base of the Cerros for several miles until it joins the Río Vítor, and the combined streams flow out of the basin in a broad gap through the Cerros.

The fact that the principal sources of the Chili lie outside of the Arequipa basin, within the main Cordillera, gives the stream a character of special significance. Ogilvie (1922, p. 90) states that the measured rainfall at Arequipa for 4 years gave a mean of only 113 millimeters, but that "Above the level of Arequipa precipitation is heavier. . . . In the upper basin of the Río Chili, which lies behind the line of high volcanic peaks, the mean rainfall is believed to be about 200 millimeters." Even this seems hardly sufficient to account for the floods in the Chili during the early months of each year. In the high mountains that form part of the upper catchment

area precipitation may be considerably greater than this figure indicates. At any rate, there are remarkable fluctuations in the volume of water in the Chili. Ogilvie (1922, p. 104) gives a graph of the flow of the river at Arequipa for the years 1901–1905. During 9 months of each year the flow is almost invisible on the chart; the average annual flow of 729,000,000 cubic meters is nearly all concentrated in the first 3 months. Thus the Chili has a regimen very different from that proper to the Arequipa Basin. The rainfall of the season of 1939 at Arequipa was said to be considerably above normal, but none of the quebradas (gulches) near Arequipa showed more than a little trickle along their floors. During a considerable part of this time a swirling flood was coming down the Chili. Standing on a bridge over the stream on one occasion, I was greatly impressed by seeing bowlders that appeared to be as much as 2 feet in diameter raised momentarily to the surface. Records show, however, that at intervals of several years sudden floods of great volume come down the Quebrada de San Lázaro (the large quebrada in the center of Plate 1, figure 1 of the preceding article).

## CORDILLERA TUFFS

The preceding paper (Fenner, 1948) described the incandescent tuff flows that at one time covered the Arequipa basin to great depths. These deposits, locally named "sillar" (pronounced seel-yár) evidently resulted from a succession of flows occurring at intervals over a prolonged period.

They are of several varieties, of which detailed descriptions have been given (Fenner, 1948). The earliest, a dazzling white rock (sillar blanco) which has been used for centuries as the chief building stone in Arequipa, was followed by flows of salmon tuff (sillar colorado), and "sillar ocioso" or useless sillar. Latest of all were certain abnormal tuffs. All these, in their physiographic relations, will be considered later.

## THE QUEBRADAS AND AREQUIPA SEDIMENTS

In most respects the Arequipa basin is a pampa area of relatively low relief, but it is trenched by a considerable number of deep quebradas of rather remarkable aspect (Pl. 2, fig. 2 of Fenner 1948). Plate 2, figure 1, of the preceding paper shows quarry workings in a quebrada a few miles northwest of Arequipa (near Cerro Colorado), where most of the building-stone for the city is now obtained.

A main road running west-southwest from Arequipa toward the Pacific port of Mollendo crosses a series of deep quebradas about 20 miles from Arequipa. The road first comes to Quebrada Río Seco, and follows the rim of this for a short distance, to where this quebrada is joined by Quebrada del Crucero to form Quebrada Añashuaico. Just beyond the junction the road descends diagonally 200 feet to the bottom of the quebrada, makes an abrupt turn and ascends the other side, travels a mile or so over the pampa, and then similarly crosses Quebrada Escalerilla. The last also joins Quebrada Añashuaico, at a lower point, and Quebrada Añashuaico runs southwesterly for several miles and enters Chili River at Uchumayo.

All these quebradas are narrow gashes cut to depths up to 400 feet in the sillar of the almost featureless pampa surface. Study of their phenomena gives information of much significance.

Where these quebradas are crossed by the highway the sillar is the pure white variety, but lower in their courses there is an abrupt change to salmon sillar, which continues to the Chili River at Uchumayo. The contact between the two sillars is concealed by talus, but from relations elsewhere, the white sillar is believed to be the older.

Large remnants of sediments occur at various places within these quebradas. At one time, such deposits had evidently filled the whole depth of the quebradas and had even spread out beyond the rims. To these the name "Arequipa sediments" is applied.

Where the road first strikes Quebrada Río Seco there is a thick body of Arequipa sediments which lie partly within the quebrada, but overtop the rim and lie along it for hundreds of feet, resting upon hard white sillar containing nests of recrystallized pumice (Pl. 1, fig. 2).

From the relations here and in many other places along these quebradas, it is evident that deep, narrow canyons were cut in the consolidated tuff deposits, that these canyons received a great mass of bowldery sediments, which not only filled them but spread out beyond their rims to a considerable thickness, and that the sediments in turn were swept out, leaving only remnants as evidence.

These Arequipa sediments are largely bowlder beds, nearly black. The bowlders are andesitic lavas, up to 2 or 3 ft. in diameter but mostly smaller. Sandy layers also are present. In the photograph the layers appear regular, but though stratification is practically horizontal, irregularities of the kind produced where current action has locally cut out beds and refilled the cut are numerous. In one place the beds enclose a large bowlder of white sillar, 5 feet by 8 feet, detached from the parent body and resting in the enclosing strata a few feet above the main sillar surface, but this is one of the very few inclusions of sillar bowlders seen in Arequipa sediments anywhere in the region, which means that the erosive action that first carved the quebradas was so violent that but little sillar waste remained. If the quebradas in their present condition were to be filled with sediment brought in by streams not extremely violent, a multitude of blocks of sillar and smaller detritus would be incorporated.

The size of the bowlders of heavy, dark lavas that compose the sediments demonstrates that torrential forces continued after the quebradas had been excavated, though the interspersed sandy layers show that fluctuations occurred. It is doubtful that any great difference in the strength of floods is indicated between the first carving out of the deep gashes, removing detritus of sillar, and the later filling with bowlder deposits of lava. Rather, the change from erosion to deposition is attributed to sudden new supplies of rock waste at the stream heads, as a consequence of glacial action on the high mountains to the north. The primary trenching that preceded the filling of the quebradas with detritus indicates a prolonged pluvial period that was a forerunner of glacial conditions.

After the period of deposition most of the lately deposited sediments were swept out. This in turn was followed by the dry conditions of the present time. Where Quebrada Añashuaico enters Chili Valley, its floor stands about 20 ft. above the river, (Fenner 1948, Pl. 3, fig. 1). Erosion of the floor of the quebrada has long ceased, leaving the floor as a perched platform. The building and the stone walls are evi-

FIGURE 1. SILLAR QUARRY IN A QUEBRADA ON THE PAMPAS
Northwest of Arequipa. Peneplaned surface of white sillar. In distance, westernmost peak of Mount Chachani and foothills.

FIGURE 2. VIEW ACROSS QUEBRADA RÍO SECO
Just above its junction with Quebrada del Crucero to form Quebrada Añashuaico. Arequipa sediments rest on the sloping wall of the quebrada (lower right) and on the peneplaned rim (top left).

FIGURE 3. TERRACES AT VARIOUS LEVELS ALONG CHILI RIVER AT UCHUMAYO
Cerros de la Caldera in background.

QUEBRADAS. RIVER TERRACES

FIGURE 1. BED OF CHILI RIVER AT AREQUIPA IN THE DRY SEASON
Bowldery sediments (Arequipa beds) at left, Mount Chachani in background, with slopes covered by mounds of detritus. (Taken by a local photographer.)

FIGURE 2. AREQUIPA SEDIMENTS RESTING ON EROSION SURFACE OF SILLAR IN QUEBRADA NACACO

FIGURE 3. BEDS OF BOWLDERS AND PUMICES
In west wall of Chili Valley at Acequia Alta.

AREQUIPA SEDIMENTS

dence that no danger of floods is anticipated. Additional evidence of cessation of floods is seen in the tall trees in the quebrada in Plate 2, figure 2, of Fenner 1948. There are also one or two dwellings on the floor of the quebrada, and cultivation of crops and fruit is carried on. Since erosion ceased in this quebrada, Chili River, rising in a region of greater precipitation back of the volcanoes, has lowered its bed here about 20 feet, but it is doubtful whether erosion along the Chili is going on at present.

The sedimentary strata (Arequipa sediments) that overtop the rims of Quebrada Añashuaico rest everywhere upon a surface that is almost an ideal plane (Pl. 1, figs. 1 and 2). Likewise, deposits of sillar ocioso (the third type of sillar) rest upon this planed surface at various places. The planation acted upon deposits of sillar blanco and sillar colorado, but before the flow of sillar ocioso and deposition of sediments.

Large quantities of talus mantle the walls and descend to the bottom of Quebrada Añashuaico in many places. Evidently no torrential floods of the sort that first excavated the quebrada have flowed down its constricted channel for many years, probably centuries.

Large remnants of Arequipa sediments occur at intervals along the whole length of Quebrada Añashuaico and its tributaries (Fenner 1948, Pl. 1, fig. 2; Pl. 2, fig. 2).

In this same region southwest of Arequipa there is another important quebrada: Quebrada del Huaico. The upper (northeastern) part of this is in white sillar, but lower down, the walls are formed of salmon sillar, as in Quebrada Añashuaico. Along Quebrada del Huaico the same planation of the two early sillars is found.

To the right of outlet of Quebrada Añashuaico (Fenner, 1948, Pl. 3, fig. 1) is the base of a hill of gneiss which represents a spur of the Cerros de la Caldera that was cut off from the main body when the Chili was forced against the flanks of the Cerros by the Cordillera tuff flows, and eroded a new channel for itself. The present course of the Chili for many miles along the flanks of the Cerros is unnatural. The river's natural position is somewhere in the broad valley of the Arequipa Basin north of here, which is now filled with tuff deposits and remnants of sediments. Instead, it hugs the Cerros and angles back and forth at each spur, or, in several instances where the spur was completely buried by the tuff, the river has cut across the hard gneiss and entrenched itself in narrow canyons (Fenner 1948, Pl. 2, fig. 3). The course along the base of the Cerros is one that it took when tuff flows forced it southward out of its original channel. Its later history was parallel to that described for Quebrada Añashuaico—carving out a new valley, filling this valley with bowlder beds, and again removing those deposits. Naturally the events that occurred in Quebrada Añashuaico required that analogous events should have occurred in the Chili valley into which Quebrada Añashuaico empties.

## PRELIMINARY OUTLINE OF SEQUENCE OF EVENTS

Anticipating a little the significance of certain features yet to be described, we may summarize the inferred succession of events in the history of these quebradas during the Pleistocene as follows:

*Stage 1. Pre-sillar conditions.* Chili River flowed over a basin toward a south-

western exit in the large gap through the Cerros de la Caldera (Fenner 1948, Fig. 1). The depth at which the old surface of the basin is buried beneath the sillars is unknown, for the present deep quebradas have not discovered it. At this time the great volcanoes at the northern and northeastern borders of the basin, Mounts Chachani, El Misti, and Pichu Pichu, are believed to have been long in existence.

*Stage 2. Flows of white sillar.* Flows of white sillar covered the basin to a great depth. Large remnants of these flows are found to a height of 8400 feet on the southwestward-facing slopes of El Misti, and a probable remnant at about 9300 feet was seen from a distance. These flows probably spilled over at low places in the Cerros de la Caldera, and a great deal must have flowed out of the basin through the large southwestern gap. In the eastern part of the basin, *sillar blanco* occurs near Chihuata at 9000 feet and near the Rescate Mine at 8300 feet. At this time, Chili River was probably displaced from its original course by the sillar flows and pushed over toward the southwestern base of El Misti, and, lower in its course, against the northern flank of the Cerros de la Caldera.

*Stage 3. Erosion of white sillar.* Great erosion of the white sillar followed in southern and western areas. Although this country must have been covered by white sillar in Stage 2, none is exposed there now in the deep quebradas. Its place has been taken by *sillar colorado*. A large area of white sillar was left in the northern and central areas, as shown by the quarries near Cerro Colorado, by the exposures along the upper courses of Quebradas Río Seco, del Crucero, Escalerilla, and Añashuaico; and exposures along the northeastern part of Quebrada del Huaico.

The extent and depth of the denudation of *sillar blanco* indicate that the period was a prolonged one. Probably the course of Chili River across spurs of the Cerros de la Caldera was first established during this period.

*Stage 4. Deposition of sillar colorado.* Flows of *sillar colorado* next filled the depressions from which white sillar had been removed, and it must also have covered deeply the whole basin (or at least the western half of it), for in the great canyon of the Chili north of Arequipa this sillar attains an elevation of more than 9,000 feet. Generally it is a moderately hard rock, but in places weathering has disintegrated it so that it is difficult to distinguish from *sillar ocioso*, except that *sillar ocioso* was deposited after the great planation of Stage 5.

*Stage 5. Planation of sillar blanco and sillar colorado.* In many areas the earlier sillars are concealed by *sillar ocioso* of the later Stage 6 or by gravel and bowlder deposits of Stage 8, but, over a wide region, wherever the two earlier sillars are exposed, as in the form of shelves jutting out from under the later deposits along the rims of quebradas, they have an astonishingly plane surface. This surface passes without interruption from *sillar blanco* to *sillar colorado*. It is an outstanding feature along Quebrada Añashuaico and its tributaries.

In Quebrada del Huaico the relations are analogous. The white sillar in its upper (northeastern) part is succeeded by *sillar colorado*, which continues southward down to Chili River. The two sillars are planed off. The same planation appears for many miles in the bluffs facing the river (Fenner, 1948, Pl. 2, fig. 3).

The planed surface has a uniform gentle descent from northeast to southwest. At the quarries near Cerro Colorado, in the middle part of the basin, its elevation is

about 2400 meters; near the town of El Huaico it is a little below 2200, and on the bluffs above Uchumayo a little above 2000.

This extraordinary planation, as well as other topographic effects, is ascribed to derangements of the drainage of Chili River by the sillars themselves. During Stage 1, Chili River escaped from the basin through the wide southwestern gap in the Cerros de la Caldera. When the flows of *sillar blanco* (Stage 2) occurred, the inference is that they filled the channel of the river through the gap (as well as all drainage lines throughout the basin). During Stage 3, either this channel through the gap was rediscovered or (more likely) a new channel became deeply entrenched, so that erosion in the basin carried away a great part of the *sillar blanco*. Then flows of *sillar colorado* (Stage 4) obliterated the exit channel for a second time. The course of the river was thereby shifted so that in entrenching itself it encountered a long stretch (perhaps several miles) of crystalline gneiss in the gap, along which erosion was extremely slow in comparison with erosion of sillar in the basin. Thus time was given for the wonderful planation of the two sillars.

Further evidence of this planation appears in a concordant terrace on the flank of the Cerros de la Caldera at Uchumayo (Pl. 1, fig. 3, background) and seems to be recognizable at the appropriate elevation on the Cerros farther up the river, also in certain features discovered by Dr. W. F. Jenks (personal communication). He has found "splendid pediments on hard rock at Vítor and northwest of the Yura River." These localities are in the vicinity of the southwestern gap, along streams of the same drainage system as the Chili. His observations seem to fit in with my deductions as to the cause of the widespread planation described above.

*Stage 6. Flows of sillar ocioso.* This is the almost powdery sillar that covers the erosion plane of Stage 5 to a considerable depth in many places. Once more the course of Chili River through the gap is believed tohave been blocked, and the channel shifted to a new position.

*Stage 7. Re-entrenchment of Chili River. Excavation of tributary quebradas.* In the new position Chili River seems to have rediscovered one of the old, deep channels through the gap and to have soon entrenched itself to its former depth. Erosion worked rapidly up-stream, and new tributaries were cut through the planed surface to a depth of several hundred feet: Quebradas del Huaico, Añashuaico, Escalerilla, del Crucero, and Río Seco., as well as others to the west that have been observed but less thoroughly examined. The character of these quebradas points to rapid entrenchment by torrential streams.

*Stage 8. Filling of the lately entrenched channels of Chili River and its tributary quebradas with bowlder and gravel deposits: Arequipa sediments.* Before these rapidly excavated channels had been widened by lateral planation, new events occurred. So far as any evidence can be seen, the Chili, at the end of the last stage, was flowing in an unobstructed course; and no new obstruction now appeared. Its bed was a few feet higher than at present and the beds of its tributaries were conformable. In spite of the lack of extraneous obstruction, all these channels became filled with deposits of bowlders, gravel, and sand. These deposits even overtopped the rims and spread over the intervening pampas to a considerable depth. They appear to be the normal result of the overloading of torrential streams of glacial origin, acting

upon deposits of morainal detritus that had suddenly become available. The evidence for this glacial origin will be fully considered in a later section. Enormous heaps of bowldery detritus, which lie at the bases of the three volcanoes are believed to be essentially morainal, but glacial action in this region was complicated by a number of factors not found in the continental glaciation of the northern hemisphere. This introduces new possibilities, and Dr. Jenks favors a different interpretation of the bowldery deposits. What I would bring out here is that whether the morainelike deposits are really of glacial origin or not, their situation and make-up are such that they would serve, like moraines, as sources of supply for overloaded streams, in a pluvial climate, to fill the quebradas of the Arequipa region with bowlders. The same sort of aggrading process that has been recognized in glaciated regions in countless instances, produced here such features of aggradation as outwash aprons, overwash plains, frontal aprons, valley trains, and so on, filling the quebradas and covering the pampas. No rise of sea level or extraneous blocking of stream flow needs to be postulated to account for it.

The aggradation that so frequently accompanied glacial action in the northern hemisphere resulted from the action of large volumes of melt-water upon accumulations of morainal material, whereby the streams became overloaded and deposited their burden as soon as the inclination lessened. In the Arequipa region the conditions were similar, whether the great piles of bowldery material at the source be morainal or not. The volume and strength of stream-flow at this time are attested by the large size of transported bowlders; nevertheless the streams were overloaded as soon as they left the steep slopes at their source.

Much of the sedimentary material deposited in the quebradas at this stage was removed in the next stage, but the amount left more than suffices to show its character and position (Pl. 1, fig. 2). Evidently times of torrential flow alternated with relatively quiet periods, and the strata show many erosional unconformities. It is difficult to form an opinion of the duration of time represented by such hiatuses. Possibly there were extended periods of erosion, but I have seen no evidence in this region of a great break, such as would imply two distinct glacial periods. Neither among the sediments themselves nor among the morainal sources does the amount of weathering imply that part of the material is definitely of an earlier age than other parts. The most definite implication of a rather prolonged interval of quiet conditions is found in certain diatomaceous lake beds of Pleistocene age at Chihuata (see later). The glacial period in this region was undoubtedly a time of many vicissitudes, both climatic and volcanologic.

*Stage 9. Removal of bowlders and gravel deposits from quebradas and from the valley of Chili River.* The sediments deposited during Stage 8 were almost wholly removed in Stage 9. This is attributed to increased flow during the final melting of the glaciers, as is common in glacial deposits of the northern hemisphere. In the Arequipa region cessation of sedimentary deposition cannot be attributed to exhaustion of supply at the source, for great piles still exist. It seems, then, that the force of floods became sufficient, at times at least, to sweep out the bowldery sediments.

At the present time Chili River, originating in a drainage basin back of the volcanoes, is subject to seasonal floods which might be thought sufficient to cause ero-

sion under present conditions, but the flood waters bring down bowlders from the upper reaches and there is no distinct evidence that either erosion or aggradation prevails. The large quebradas southwest of Arequipa are certainly not being eroded, and it is doubtful that erosion is proceeding in any of the quebradas that come down the southwestern slopes of El Misti. In Stage 9, however, the sediments in these quebradas were carried away and similar deposits on the nearly flat pampas were removed. The entrenchment of the quebradas, their filling with sediments, and finally the removal of the sediments are all believed to be evidences of the prevalence of a pluvial climate during the Pleistocene very unlike the aridity of today. As will be shown later, this conclusion finds support in the views of Bowman, (1916) who speaks repeatedly of the wetter climate of the Pleistocene as the cause of valley filling.

*Stage 10. Late terracing along Chili River.* Several distinct terraces of very modern aspect are observable at moderate heghts above the river. The village of Uchumayo stands on a broad platform 60–75 feet above the river (Pl. 1, fig. 3, foreground). A lower terrace appears in the middle distance. (The high, level bench in the background is probably much older. It corresponds in elevation to the planed surface of the sillars of Stage 5.)

The low terraces along Chili River indicate late climatic fluctuations. At first sight, slowing-down of river erosion by the out-jutting spurs of the Cerros de la Caldera might seem a likely cause, but the spurs were probably cut through much earlier than the formation of the terraces, and although erosion of the hard gneis would slow down the process of erosion it would not cause a definite halt at any stages

## RELATIONS OF SILLARS AND AREQUIPA SEDIMENTS ALONG THE CHILI VALLEY NORTH OF AREQUIPA

Certain events in the late climatic and topographic history of the region west and southwest of Arequipa have been described. In the region north of the city, topographic results should be expected to be similar but not identical, because this area embraces the lower slopes of Mount Misti and the adjacent valley of Chili River.

Plate 4, figure 1 of the preceding article shows terraces and graded slopes on the east side of Chili Valley north of Arequipa. On the left are the lowermost slopes of the southwest side of Mount Misti. They rise from here on a gradually increasing grade to the summit of the mountain. The material underlying the slopes in this view is sillar in one place, Arequipa sediments in another. The slopes are by no means original surfaces of deposition of either volcanic or sedimentary material, but have been graded as a unit at a late period. Here and in other places I find reason to differ with a statement by Ogilvie (1922, p. 21) that "speaking generally the slopes in the Western Cordillera are more or less the original slopes of deposit, whether of the lava flows or of the agglomerates and ash beds which have fallen around some center of eruption." As an example Ogilvie pictures almost the identical scene pictured here. In my view there is hardly a surface to be found in the Arequipa region, whether of mountain slopes, of moraines, of pampas, or of river terraces, that is not either wholly the result of erosion or has not been fundamentally modified by it.

The relations of sillars and sediments in this area are analogous to those described

for the pampa region to the southwest, but there is one marked difference: the planation of sillar deposits of Stage 5, so prominent on the pampas, does not appear. There are deep quebradas in sillar, descending the mountain slopes into the wide valley of the Chili, and these have the same history of incision, filling with sediments, and removal of the filling, but the areas between large quebradas are cut up by smaller channels, and the surface of the sillars is very irregular.

A reasonable supposition is that planation of sillars during Stage 5 was effective here as elsewhere, but that it could not advance far eastward into the mountain mass of El Misti. Instead, the plane merged into a gently inclined pediment running back up the adjacent slopes of the mountain. During the erosion of the subsequent period, Chili River, then a powerful stream in the pluvial climate destroyed the planed surface along its course and eroded a wide valley. At the same time the pediment was cut up into quebradas and lesser ravines by streams descending the contiguous slopes of Mount Misti. When glacial conditions on the mountains supervened, sedimentary detritus was deposited in both the Chili Valley and the side quebradas. In more recent times, there were attempts to form new pediments at successively lower levels on the mountain slopes to meet successively deeper entrenchments of Chili River. These pediments beveled sediments and sillars alike. In the final stages, most of the sediments were cleared out and low terraces along Chili River were formed, as depicted in the photograph.

Where Chili River flows through Arequipa its banks are bowlder and gravel beds, firmly cemented, (Pl. 2, fig. 1). A few miles to the north a thick series of sediments appears on the west side (Fenner, 1948, Pl. 4, fig. 2). On the left side of the photograph sillar appears under the sediments, and its surface rises toward the left until (outside of the view) it reaches the top of the valley wall. It is on this mass of sillar that the old Harvard Observatory stands.

The sedimentary beds are composed of firmly consolidated bowlders, gravel, and sand, with intercalated strata of pumice. The beds are approximately horizontal but quite irregular, with members abruptly cut out. They have the characteristic appearance of the Arequipa beds in many other places; that is, they are torrential stream deposits alternating with finer layers. The components in general are dark andesites, but there are layers of pure pumice or pumice mixed with andesitic gravel. It was difficult to obtain a representative photograph of the strata here, for the narrow burro trail that descends the bluffs lies too close to the face. Figure 3 of Plate 2 includes two beds of pumice, one just under the point of the pick and a thinner bed a little above. Such pumice beds may be a foot or more thick, extend for 100 feet or more, and sometimes split and re-unite. Much of the pumice is white and apparently rhyolitic, but some contains phenocrysts of pyroxene, hornblende, and feldspar in a dark groundmass. Maximum size of lumps is about that of a hickory nut; much of it is rounded. Most of these pumice beds look like water-laid material, but some of them may have been contemporaneous deposits of ash-falls. Conglomeratic beds contain lava bowlders 2 feet or rarely 4 feet in diameter that are either somewhat angular or rounded. A considerable amount of ground-up pumice and sillar is included. A rough guess put the total amount at 10 per cent by volume.

The character of the series is that of a succession of strata laid down by violent

FIGURE 1. GREAT CANYON OF RÍO CHILI NORTH OF AREQUIPA.
Walls of salmon sillar about 500 ft. high. At upper left is cap of Chili andesite.

FIGURE 2. REMNANTS OF WHITE SILLAR
Sillar had filled a narrow gash about 250 ft. deep in San Lázaro andesite, Quebrada Huarangueros, at 8400 ft. elevation on the slope of El Misti. City of Arequipa appears in distance.

FIGURE 3. QUEBRADA DE SAN LÁZARO
Stages of erosion of unconsolidated deposits at base of Mount Misti. Note planed-off remnants within the quebrada, just above center.

CANYONS CUT IN SILLARS, IN ANDESITE, AND IN MORAINAL MATERIAL

FIGURE 1. MORAINAL HILL AT YUMINA

FIGURE 2. HETEROGENEOUS BOWLDER DEPOSITS AT SOUTHERN BASE OF MOUNT CHACHANI
Material mostly a loose aggregate, but solid rock appears at upper left

FIGURE 3. MORAINAL MATERIAL ON SOUTHERN SLOPES OF EL MISTI
Near Cuesta de San Bernardino.

GLACIAL DEPOSITS

torrent action alternating with quiet stream flow. The numerous beds of pumice indicate frequent explosive volcanic activity somewhere in the immediately adjacent region. The sediments shown in the figures are to be referred to the same period and the same conditions of formation as those studied in Quebrada Añashuaico and other quebradas southwest of Arequipa. They are regarded as having been derived from morainal material carried down the slopes of the great volcanoes and distributed, together with contemporaneous pumiceous ejecta and ground-up sillar, along outwash channels of streams by floods and more quiet flow.

On the opposite (eastern) side of Chili valley the long, graded slopes at the base of El Misti (Fenner 1948, Pl. 4, fig. 1) are underlain in places with white sillar, elsewhere with salmon sillar. Overlying the irregularly dissected surfaces of sillar are large or small areas of Arequipa sediments.

Several large quebradas and numerous small channels that were cut into the sillars in this area were filled with these sediments, and have been re-excavated (Pl. 2, fig. 2). In this quebrada (Nacaco) salmon sillar in the walls rises to a height of 50 feet or more, or drops below the visible sediments. In some occurrences thinly layered sandy sediments are evidently composed of ground-up pumice and sillar with little foreign material, and doubtless all the sediments contain much pumice and sillar. Some of the white sillar in Quebrada Nacaco is of rocklike hardness and, superficially, looks like a lava flow.

Plate 4, figure 3, of Fenner, 1948, shows the occurrence in Quebrada Nacaco that was described in the previous paper, where, during the general period of deposition of Arequipa sediments a thick section was washed out, and the gap was later filled with a jumble of heterogeneous material, and this in turn was covered with sediments.

One of the highest exposures of sillar in this territory is shown in Plate 3, figure 2. Here a narrow gash had been formed by some condition of rejuvenated stream erosion in andesite country rock at a period prior to the sillar extrusions. The sillar attained sufficient depth at the mouth of the cut to enable it to flow back into the cut, but most of this filling has been removed.

The andesite that forms the country rock here, called San Lázaro andesite, is itself of much interest. It is evidently a very old rock, which forms a large, thick body on the slopes of Mount Misti. Farther up the mountain, at an elevation of about 9,100 feet, an almost flat surface of several square miles has been developed on it. This platform seems to have no reason for existence except as the remnant of an old peneplain, on which the volcano has been built up. Other exposures of San Lázaro andesite extend as an almost continuous scarp, almost buried in sillar, for several miles southeasterly, to the Quebrada de San Lázaro.

North of the region just described, the river emerges from a profound canyon between Mount Misti and Mount Chachani. In the interval the valley gradually narrows. On the west side the wall of Arequipa sediments underlain by sillar continues. The hummocky surface of sillar rises and the overlying sediments become less prominent. The river flows in an inner trough, 50 feet or more below the general valley floor, with walls of white sillar.

On the east side the valley is bounded by a lava flow overlying salmon sillar and gravel. This lava cap, called the Chili andesite, is a late feature. It undoubtedly

followed a former river channel when the channel stood at a considerably higher level than at present. The lava is a dark rock, with laminations dipping in various directions and with prominent protoclastic structure. The underlying sillar has an uneven surface, and in one or two places rises to the base of the lava cap, but usually gravel beds lie between.

This lava continues northward as a continuous sheet, capping the old channel gravels on the east side of the valley (Pl. 3, fig. 1; Fenner, 1948, Pl. 5, fig. 2) and thence into the canyon. Farther up the canyon the river has cut through the lava sheet, and it appears at its appropriate level on the northwest side and continues upstream. The last of it was seen at an elevation of about 9,000 feet, as the edge of a thin layer, the principal part presumably having been cut out by the canyon. Gravel beds lie above and below it. In most places in the canyon the lava rests on thick beds of gravel, with salmon sillar below.

The walls of Chili Canyon are almost wholly of salmon sillar. Evidently several thick flows occurred, building up piles estimated to be 500 feet thick in their present exposures.

The characteristics of the Chili andesite that flowed down the old river channel and its relations to other features seem to outline the following history for the canyon: At an early period the river had flowed in a narrow gorge between Chachani and El Misti, at about the elevation of the present channel. Into this gorge hundreds of feet of tuff flows were piled. On top of these the river resumed its course. At first it was an eroding stream which, after leaving the canyon and coming out into the open country north of the site of Arequipa, wandered over the tuff in a wide valley. Then, probably for a climatic reason, it became an aggrading stream of small velocity and deposited thick gravel beds. Next, the flow of Chili andesite down the channel formed a narrow tongue within the canyon and spread out in the wider country below the exit. The river resumed its flow, and for a time deposited gravels on top of the lava, but it shifted its course to the softer tuffs at the side, and has worked down to about its old grade.

The canyon is undoubtedly a very old feature, which provides the outlet for a large drainage area to the north (Fenner, 1948, Fig. 1). Within the canyon the river is narrowly confined by the great bulk of Mount Chachani on one side and El Misti on the other. For ages these have been shedding their debris into the channel, and only by the flow of the river has the channel been kept open. Mount Misti looks like a young volcano, and geologically it doubtless is, but it is probably much older than the first sillar flows here.

## UNSORTED DETRITAL DEPOSITS

The tentative assumption has been made that glaciation on the high mountains of the region was of great importance. On the slopes and along the bases of these mountains and, in places, out on the pampas at some distance from them, are great heaps of unsorted detritus, which have been assumed here to be of morainal origin. Whether they were actually morainal or are of other derivation, they served as sources of the material that was swept by torrential streams into valleys and quebradas and over pampa surfaces, and deposited as the Arequipa sediments.

In studying these bowlder heaps in the field I had no doubt that they were essentially glacial deposits, and in subsequently reviewing the evidence and considering other possibilities no explanation that does not recognize glaciers as the chief agency has appeared to me satisfactory. On the other hand, W. F. Jenks, studying some of the same phenomena, has reached different conclusions. He believes (personal communication) that glacial action in the Andean region of southern Peru was confined to the higher elevations. He sets 14,000 feet as the common lower limit but recognizes that some glaciers may have descended to 12,800 ft. (3,900 meters), and possibly in rare instances to 10,980 feet (3,300 meters). The accumulations of detritus at low elevations around the volcanoes of the Arequipa region are ascribed by Jenks to volcanic mud flows or lahars.

Pleistocene conditions here, both regional and local, were different in several important respects from those prevailing in the northern hemisphere in regions of continental glaciation, and an endeavor must be made to evaluate the effect of these conditions. One must consider whether the observed results might not have been due, as Jenks concludes, to causes other than glaciation. Correspondence between us has not enabled us to reconcile entirely our different views; therefore each will present the basis of his judgment in a separate article.

The Arequipa basin is only a little more than 16 degrees south of the equator. On its northern and northeastern sides three great volcanoes rise as a wall to heights of 18,583 to 19,931 feet, from bases at about 8,000 feet. The steepness of their slopes would greatly facilitate the downward movement of glaciers, so that masses of ice, formed in the upper frigid regions, would have a relatively brief journey through the lower zones of ablation. This would favor their descent to regions well below those that would be reached on gentler grades.

At the same time, these steep slopes would favor the action of other, radically different processes that might account for the transfer of clastic material from the higher slopes and its accumulation below. As a basis for discussing what these other processes might have been, a description of the character and the situation of the heaps of detritus in relation to the mountains will be given.

Northwest of Arequipa the basin is dominated by the great bulk of Mount Chachani. Along its southerly base are rounded heaps of unsorted, bowldery detritus (Pl. 4, fig. 2). In places this detritus coveres rock *in situ* but generally its unconsolidated character is shown by ravines and gullies that have readily cut back into it.

Dumps such as these, here and elsewhere, demonstrate that deposition of Arequipa sediments ceased, not because supplies of easily available and transportable material are not at hand, but because the floods that had acted as the transporting agent have ceased. The originally great bulk of the Arequipa sediments, though only small remnants now exist, show how greatly the detrital dumps must have been eroded to supply material.

Covering the slopes of Mount Chachani to a great height, similar heaps of detritus conceal the rocky frame of the mountain (Pl. 2, fig. 1). The mounds are dissected by large and small ravines, rarely revealing outcrops of rock. These slopes are almost obscured photographically by a stratum of light-blue haze which is almost opaque to actinic rays, though visually details are clear.

The uppermost parts of the mountain consist of several main peaks with minor subdivisions (Pl. 2, fig. 1; Fenner 1948, Pl. 1, fig. 2). Probably everyone who has studied them would agree that the highest summits of Mount Chachani have been profoundly acted upon by glaciers. The easternmost of the main peaks (at right of views) has been carved into a cirque, but no glacial ice can now be detected on it. The trough between this peak and the next on the west has also been modified by glacial action. The rocky ridges of the summit extend down the slopes to where the detrital heaps begin, and these heaps apparently represent the sites of deposition of the glaciers.

A little to the north of Arequipa, on the southern edge of the Pampa Polanco, smaller, somewhat isolated heaps of unsorted, rounded bowlders cover a considerable area at a much lower elevation. They lie definitely out in the pampa region at the base of El Misti, but the lowest slopes of the mountain are not far away.

In the airplane view of El Misti (Fenner, 1948, Pl. 1, fig. 1), the large quebrada in the center is Quebrada de San Lázaro. On the left of the entrance is a large exposure of San Lázaro andesite. On the right is a heap of unconsolidated detritus, which rises abruptly from the pampa to a height of about 200 feet, and continues up along the side of the quebrada as a boundary wall. The component material looks like typical moraine, filled with bowlders. One of these was estimated to be 15 ft. in diameter. The rim of the bounding wall has a nearly plane top that slopes upward toward the summit of the mountain at a gentle grade, probably 3 or 4 degrees.

This wall is only an intermediate boundary of the quebrada. A comprehensive view of the surrounding country from a height of 9100 feet on the rim showed other graded surfaces, rising in steps parallel to the quebrada, several hundred feet higher, at a little distance back from the rim. These likewise are gently inclined upward toward the summit of the mountain, but at their lower ends they are cut off abruptly where they meet the pampa. At places in Quebrada San Lázaro (Pl. 3, fig. 3) the walls on the right are capped with a thick layer of rounded bowlders and stratified sediments, dipping down-quebrada, which looks like a stream deposit.

Most of the material of the walls appears morainal, but in places it is a mixture of pumiceous tuff and rock fragments, clearly not morainal but resembling certain abnormal tuff deposits described in the preceding article.

The depositional history of the detritus that forms the walls of this quebrada and covers the mountain slopes hereabouts is evidently complex, and not all the relations have been worked out, but moraines and abnormal tuffs are believed to be involved. Erosional history is likewise complex. None of the sloping surfaces appear to be surfaces of original deposition, but rather to be segments of former mountain profiles, the result of successive efforts to form pediments graded toward a succession of temporary base levels, out on the pampas beyond the mountain.

Within the quebrada, further (later) stages of planation show temporary halts in erosion, as illustrated by the remnants (Pl. 3, fig. 3). Similar remnants are visible from here far up the quebrada, with their inclination becoming greater at higher levels.

These features show that all surfaces here have been modified by erosion. We cannot look for the typical outward forms that are sometimes thought to be the necessary

FIGURE 1. MORAINAL HILLS WHOSE SECONDARILY GRADED SURFACES SLOPE AWAY FROM MOUNT MISTI
Near Cuesta de San Bernardino.

FIGURE 2. DIATOMACEOUS SILTS OF VARVELIKE STRUCTURE AT CHIHUATA

BOWLDERY ACCUMULATIONS. DIATOMACEOUS SILTS

result of morainal deposition. If such were originally present, erosion has modified them beyond recognition.

Quebrada de San Lázaro lies a few miles northeast of Arequipa. A greater development of moraines lies in a region nearly due east of the city and south of Quebrada de San Lázaro, on the road to Chihuata. This town is about 10 miles in an air line from Arequipa.

The great bowlder heaps on the road to Chihuata form a wall plainly visible from Arequipa. Its westward-facing slopes lie 4 or 5 miles away.

A road ascends them, winding into gullies and out around spurs, over unsorted debris exposed in numerous road cuts. After crossing the summit at Cuesta de San Bernardino (San Bernardino grade), at an elevation of about 9,300 feet the road descends slightly to Chihuata (9,000 feet).

Over a wide area near Cuesta de San Bernardino (Pl. 5, fig. 1) no solid rock is visible. Quebradas cut down for several hundred feet, and their bottoms appear to be in the same detritus. This topography extends far up the slopes of Mount Misti, whose summit lies about due north from the Cuesta de San Bernardino. Southeastward the detrital rampart continues for a long distance in an arc along the southwestern base of Mount Pichu Pichu.

On these slopes of El Misti the surfaces have been graded to a pediment, adjusted to accord with the broadly sweeping downward curves of the mountains. The profile of El Misti (Pl. 5, fig. 1) is a fine example of the manner in which erosion, acting upon easily transportable material on a conical volcano results in a smooth curve that gradually decreases in inclination from the summit to the base. The later incision of deep gulches to meet a lower base-level of erosion hardly mars the symmetry.

In the vicinity of the Cuesta de San Bernardino, the detritus varies in the size and the relative number of bowlders. In places it is sandy or even faintly stratified. The rocks are predominantly medium andesites, with many hornblende phenocrysts (Pl. 4, fig. 3). Beds of pumice, a few inches to a foot or more thick, occur near the surface but not in the masses below. The degree of rounding of bowlders varies greatly. Rounding is not prominent in the photograph, but this material has not been carried more than a few miles at most, down the slopes of Mount Misti, and under such circumstances angularity of transported bowlders is not unusual in moraines. Striking examples of such angularity under similar conditions are illustrated in photographs accompanying Louise A. Boyd's "The Fiord Region of East Greeland" (1935). In many examples there pictured the rocks in the moraines at the foot of ice-walls of living glaciers appear almost as little rounded as if recently blasted from rock-cliffs. The rounding of bowlders in deposits of this kind is supporting evidence of glacial action, but lack of rounding is not contradictory.

In an article accompanying Boyd's report (1935, p. 161) Bretz speaks of remarkable instances where morainal bowlders, now lying in the beds of torrential streams or in solifluction trains, preserve their angularity.

At the same time, pyroclastic eruptions and lava flows in the Arequipa region may have contributed to the material that the glaciers were transporting, and rocks from such sources, perhaps carried on the backs of glaciers, might be little modified by glacial grinding.

Near Chihuata, white sillar appears along the valleys of small streams, and there are large exposures of extraordinary silty sediments. Most of the latter are in horizontal beds, but in one occurrence the strata are steeply inclined (Pl. 5, fig. 2). There is no cross-bedding here, but the strata are slightly wavy. The photograph exaggerates this waviness because of perspective in looking at an irregular face. Hundreds (or possibly thousands) of alternations of extremely fine silt with slightly coarser sand suggest annual varves. The strata as a whole look like foreset beds in the delta of a lake. If such near-shore deposition was their origin, it seems to mean that the stream entering the lake did not experience torrential flow for a very long time. The appearance of the beds suggested diatomaceous deposits. A sample was taken, and later microscopic examination revealed many diatoms, mixed with feldspar grains and shards of glass.

The sample was submitted to Professor H. N. Coryell of the Department of Geology of Columbia University, to whom thanks are due for the determination of the following species of fresh-water diatoms represented:

*Stephanodiscus astraea*—disc form, medium size,—.05 mm.d.
*Stauroneis phoenicenteron*—straight spindle, large form—.11 mm.l.
*Pinnularia viridis*—elongate with parallel sides, large form—.09 mm.l.
*Pinnularia major*—elongate with parallel sides, very large form—.21 mm.l.
*Cyclotella striata*—disc form—.03 mm.d.
*Campylodiscus echeneis*—disc form, large form—.14 mm.d.
*Surirella* cf. *oblonga*—large elongate form—.18 mm.l.
*Stenoplerobia* sp.—very long and slender.

Professor Coryell states:

"The above list is a portion of the diatom forms that are found in the deposit. These are most common in the Pleistocene deposits of the Gulf Coast of North America. Other genera not fully studied include Cymbella, Fragelaria, Navicula, Denticula, Epithernia, and Hyalodiscus. . . .These diatoms are mostly from warm to average temperate zone climates."

These silty beds should be more fully investigated. They are very unlike the heavy bowlder deposits that characterize Pleistocene sedimentation elsewhere in the region. They seem to imply a rather long interval of equable precipitation at some stage of the Pleistocene. Search should be made for other forms of life than the diatoms.

The hills of morainelike detritus in the vicinity of Cuesta de San Bernardino (Pl. 5, fig. 1) extend several miles southeasterly and then easterly in an arc around the base of Mount Pichu Pichu. They have not been closely examined in this area, but the topography of their front, as seen from a short distance, appears of the same form as that of the Cuesta de San Bernardino portion. It seems likely that they are composed of morainal detritus that originated in Mount Pichu Pichu.

Certain other deposits in this territory southeast of Arequipa, rising as low hills above the pampas, are of special interest because of their isolated position, several miles from the base of the mountains.

The first are at Paucarpata and near-by Sabandía. The material here is a strongly cemented, heterogeneous mixture of bowlders, large and small, gravel, and sand, wholly unsorted. The largest bowlders are 8 to 10 feet in diameter and most are well rounded.

A couple of miles eastward of Paucarpata is a small hill, on which is the little village of Yumina. Here the surface is covered with large bowlders (Pl. 4, fig. 1). In the distance, looking easterly, is the range of high hills running southeasterly from the bowlder deposits of the Cuesta de San Bernardino. Strung across the pampa between this range and Yumina is a succession of small, squat hills, very like that at Yumina.

## POSSIBLE METHODS OF ACCUMULATION OF UNSORTED DEBRIS

There are various processes by which masses of bowlders, more or less mixed with fines, might originate and be moved down mountain slopes, and accumulate in piles at the bottom. These may be listed as follows:

Glacial movement.
Volcanic discharges, pyroclastic or as lavas.
Stream transportation.
Solifluction (Rock streams).
Ash slides.
Lahars.
Mud flows.
Rock avalanches. } Especially as touched off by earthquakes.
Rock slumps.

Some of these processes are likely so to merge with others that strict lines of separation cannot be drawn, or two or more processes in sequence may be operative. In the Arequipa region effects of climate during the Pleistocene were so combined with those due to a steep mountain topography and probably with volcanic and seismologic disturbances that one would hesitate to exclude any of the processes listed.

Rock avalanches or rock slumps may have occurred here, but hardly as a chief cause. The chief feature of several immense slides of rocks or mixtures of rocks and mud whose effects I have seen in Alaska and Yellowstone Park was their demonstration of violence of movement and tendency to become dispersed. Even after reaching nearly level ground they continued like a flood of water and did not build up great piles or mounds. The same feature is prominent in descriptions of famous slides elsewhere, such as the Gohna slide in the Himalayas, that of Bandai San in Japan, that at Elm, Switzerland, and that at Frank, Alberta. Those who have described them have been amazed at their remarkable fluidity.

Undoubtedly, less violent slumps of rock or of soil and rock occur, and if on nearly flat ground and if they have not descended from a height, they may result in large mounds; but the bowlders contained in them are not rounded unless derived from previous deposits of worn bowlders. Flows of very fluid mud and rock streams (solifluction effects) also may have been of some importance, but I know of no description that implies that any such manifestation has built up far-extended, huge piles such as those on the slopes and at the bases of Chachani, El Misti, and Pichu Pichu.

The possibility is recognized (and, indeed, favored) that volcanic action, in the form of pyroclastic discharges or lava flows, was an important source of material. The Pleistocene was evidently a time of great volcanic activity here. In the early

Pleistocene, before the advent of glaciers, there were great discharges of sillars, with which pumices and fragmental andesites were mingled. Then, during the deposition of the Arequipa sediments (which are outwash deposits believed to have been derived chiefly from contemporaneous moraines), ejection of sillars and pumices continued. At least two large lava flows reached the base of the mountains at about this time. One was the Chili andesite, which came down the canyon of Chili River, and the other was the large flow of andesite depicted in figure 1 of Plate 5 of the preceding article. This figure shows the terminus of a lava-stream that originated near the western end of Mount Chachani and flowed out on the pampas along a valley in sillar. The walls of the valley have since disappeared, leaving the lava flow in relief. The highway from Arequipa to Yura crosses it nearer its source, and for many miles the road winds in and out over its lobes and tongues. Jumbled masses of enormous angular blocks and solid rock *in situ* are exposed. If lavas of this protoclastic structure had been encountered by glaciers on the high slopes of the mountains they would have been easily torn to pieces and incorporated in the ice.

There is a probability, also, of pyroclastic discharges of solid rocks, such as were prominent elements of Tertiary volcanism in the western United States. Such outbursts, as well as outbursts of pumice, would cover the glaciers with insulating blankets and favor the descent of the ice to a low elevation before melting. Among the Arequipa sediments there is direct evidence of many contemporaneous pumice-falls. Evidences of the eruption of denser rocks falling upon the mountain slopes would be hard to recognize, and such contributions can only be surmised, as in accord with what we believe to have been the conditions here. Whatever the amount from such contemporaneous sources may have been (and I suspect that it was large), most of it was probably mingled in the final deposits with material of earlier origin.

The features of the deposits, as described in previous pages, are believed to point to glaciation as the chief agent, with volcanic discharges, rock avalanches, rock slumps, mud flows, and stream transportation as probable or possible contributors.

In summary, the supposition that other processes than glaciation were involved is based partly on fairly direct evidence but largely on what seems likely to have occurred. Glaciation seems to have been so predominant as to have wiped out or blurred the evidence of other processes. The view taken here is that whatever processes may have participated from time to time in supplying material, it was glacial action that gathered it up, worked it over, reassembled it, and left it in the condition and in the places where it is now found. The outstanding characteristics of the bowlder heaps are their great bulk, the manner in which they are piled up in heterogeneous and unsorted masses, their deposition on the mountain slopes right up to the rugged peaks and also, in a few instances, well out on the pampas, and the large size and complete or partial rounding of the bowlders. Taken together, these characteristics seem inconsistent with the view that any other process than the work of glaciers was the principal agent.

The manner in which the morainal material is now distributed on the mountain sides suggests that there were minor oscillations during the later stages of glaciation, by which earlier deposits were overridden, but that the general process was to lay down new deposits in succession up the slopes as the front of the glaciers retreated. As areas were freed of ice the lately deposited waste was modified by erosion.

## OBSERVATIONS ON GLACIAL GEOLOGY OF ANDEAN REGIONS BY OTHERS

Isaiah Bowman (1916) records many observations on the work of glaciers in Andean regions. Much of his work was in the eastern Cordillera and cannot be applied specifically to the western mountains. The chief value for the present purpose lies in his general conclusions regarding the climate of the Pleistocene in these regions and his recognition of glacial moraines as the source of deposits of detritus in valley fills.

He points out (1916, p. 200) that his conclusions differ from those of certain previous observers, who had contended that glacial development in the Andes of the equatorial zones during the Pleistocene was little greater than at present, and he gives reasons in detail for the contrary opinion. The further suggestion may be made here, that the special conditions discussed in the present paper as obscuring or modifying the results of glacial action may have been a factor in the failure of various observers to recognize them.

Bowman states (p. 204):

"In many places in eastern Bolivia and Peru the glaciers of the Ice Age were from 5 to 10 miles long. ... In the Juntas valley in eastern Bolivia the tree line is at 10,000 feet..., but the terminal moraines lie several thousand feet lower. In eastern Peru the glaciers in many places extended down nearly to the tree line and in a few places well below it."

In connection with the glaciation of the Cordillera Vilcapampa in the eastern range of Peru, Bowman writes (p. 205):

"The importance of the glacial features... developed on a great scale in very low latitudes in the southern hemisphere is twofold: first, it bears on the still unsettled problem of the universality of a colder climate in the Pleistocene, and second, it supplies additional data on the relative depression of the snowline in glacial times in the tropics."

At Choquetira, in this Cordillera, he found three prominent stages in the retreat of the ice. The lowest moraine, at 11,500 feet, is well dissected; the second, at 12,000 feet, is ravined and broken but topographically distinct; the third, at 12,800, is sharp-crested and regular. It is clear that "everwhere the ice retreated and advanced periodically;" and "At Pampaconas on the east the lowest terminal moraine is at least a thousand feet below timber line" (p. 211).

"In place of the twenty distinct glaciers now lying between the pass and the terminal moraine below Choquetira, there was in glacial times one great glacier with twenty minor distributaries.... Some of the glaciers were over a thousand feet thick.... Such a remarkably complete set of glacial features only 700 miles from the equator is striking evidence ... of the universal change in climate in the glacial period" (p. 214).

In the western Coast Range,

"At the head of the broadly opened Majes valley there was deposited a huge mass of extremely coarse waste several hundred feet deep and several miles long. Forward from it, interstratified with its outer margin, and continuing the same alluvial grade, is a still greater mass of finer material which descends to lower levels.... It is confidently assigned to the Pleistocene, since this is definitely known to have been a time of greater precipitation and waste removal on the mountains, and deposition on the plains and the floors of mountain valleys. Such a conclusion appears on general grounds, to be but a shade less reliable than if we were able to find in the upper Majes valley, as in so many other Andean valleys, similar alluvial deposits interlocked with glacial moraines and valley trains." (p. 230. *Note* the accompany photographs No. 155 and 156.)

"The head of the Majes valley contains at least six hundred feet and probably as much as eight hundred feet of alluvium now in process of dissection, whose coarse texture and position indicates an origin under glacial conditions." (p. 273).

"The work of the glaciers was not confined to the lofty situations. Mountain debris was delivered

to all the streams, many of which aggraded their floors to a depth of several hundred feet." (Caption of Fig. 43).

Thus Bowman describes, first, evidences of the immensely greater development of glaciers in the Andes during the Pleistocene, and, second, the consequences resulting from the overloading of streams with glacial waste and the resulting aggradation of valleys.

Lieutenant George R. Johnson, who was chief photographer of the Puervian Naval Air Service, 1928 to 1930, has published (1930) a series of excellent photographs taken from the air that include many of the mountains of this region. He directs attention to the evidences of glacial erosion. Photographs of the Nudo de Ampato, 20860 feet high, a few miles northwest of Mount Chachani, are especially interesting. The caption of his Figure 134 describes this mountain as being the only one of the volcanic mountains of this vicinity that still bears "remnants of its once great glaciers", and notes "the two large lateral moraines... with a glacial stream winding down the shallow glacial trough between them." These moraines lie on the smoothly molded, waste-covered slopes below the glacially sculptured summit regions and appear to merge with the slopes. In his Figure 136 (of Ampato) "The two large cirques at the right ... appear to be still occupied by glaciers." His Figure 129 exhibits "A well glaciated volcanic peak a few miles east of El Misti displaying remarkably fresh-looking features of glacial erosion. The three parallel cirques, ... have been cut back ... until the walls that separate them have been worn to narrow ridges." His Figure 130, of Mount Pichu Pichu, shows a glacial trough, backed by a large cirque, which "exhibits the familiar forms of glacial erosion—lateral and terminal moraines and glacial pockets." Figures 132 and 133 show glacial effects on Mount Chachani. The caption of Figure 133 reads "Chachani is scalloped by deep cirques from which the glaciers have now disappeared ... and is surrounded by the characteristic features of mountain glaciation—lateral and terminal moraines, U-shaped glacial troughs, hanging valleys." Johnson's aerial photographs show much more clearly than photographs taken from the ground the glacial features of the mountain summits. Especially noteworthy is the fact that the rugged glacial sculpturing of the summits extends right down to where it disappears beneath the waste-covered intermediate and lower slopes. Although these slopes do not present quite the forms that we are likely to look upon as most typical of morainal deposits, one can hardly conceive what disposition could have been made of the glacial detritus carried down from above unless it was deposited there. This subdued topography continues to the base of the mountains.

## FURTHER INFERENCES REGARDING PLEISTOCENE CLIMATE IN THE AREQUIPA REGION

The Arequipa Basin lies on the southwestern side of a chain of immense volcanic peaks. Just northeast of these lies the high mountainous country of the Western Cordillera. The volcanic chain is of such altitude that it might well be a climatic barrier. What were the conditions on the far side of this range during the Pleistocene?

This country is rather inaccessible, and my investigations have not been carried into it, but a certain amount of information is available from other sources.

Johnson (1930, p. 128) has published an airplane photograph of the northeast side of El Misti. The contrast of the topography of the mountain on this side with that of the southwest side (Fenner, 1948, Pl. 1, fig. 1) is striking. In comparison with the deep quebradas of the lower slopes and the furrows and ridges of the summit on the southwest side, the northeast side appears almost featureless. The only erosion visible on the mountain or at its base is a gorge near the bottom of the view, probably that of Chili River. It has been shown that the deep quebradas that are so prominent on the southwest side are not of recent origin. Their primary erosion occurred during the pluvial period that preceded the glacial. In the glacial period they were filled with waste, which was of glacial derivation, but had been re-transported and was deposited as stratified sediments. At the close of this period most of this waste of secondary deposition was removed, and since then little modification has occurred. The northeast side of the mountain seems to have been almost unaffected during this long period. Both pluvial and glacial erosion and deposition seem to have passed it by. Johnson's caption of his photograph calls attention to this contrast.

"Note the smoothness of the east side [actually the northeast side] of the cone and its lack of snow as compared with the deeply eroded flanks and snow-capped summit of the west [southwest] side. It appears . . . that the greater part of the precipitation on El Misti comes from the Pacific."

It is evident from the present studies that this contrast in precipitation prevailed during the long glacial and preceding pluvial periods.

Rondón (1937, p. 53) has given a detailed account of an ascent of El Misti, made by a group of teachers and students of the University of Ariquipa, on Oct. 31, 1935. They remained on the summit 3 days and 2 nights and made observations upon the different meteorological and topographic conditions on the two sides of the volcano.

The account of the country surveyed from the summit reads:

"Approaching the slopes that look toward Arequipa, that is, toward the southwest, enormous quebradas are noted, one of which gives birth to the 'Iloclla' [ravine] of San Lázaro, which passes through Arequipa. . . . On the northeast and north side . . . the volcano has a more natural and less dissected declivity, presenting a regular conical form, interrupted at long distances by smooth ridges that give the impression of torrents of liquid lava that precipitated themselves down the volcano. This side is constituted in great part of sand, in some places of an intense black color. . . . The vegetation on this side is scarce, some examples of Bolax Globaria (yareta), others of stipa-ichu (paja de puna). . . . This poor vegetation of the northeast and north side contrasts with the exuberance of the southwest and south . . . where a great variety of plants abounds . . .

"The greater part of the year the summit of this volcano is covered with snow, varying in quantity at different seasons of the year. Very few are the days that there is no snow . . . It is common to see a snow-covering down to 800 meters below the crater. On other occasions the snow covers the whole volcano completely. Ordinarily the collection of snow is greater on the southwest and south side, . . . less toward the north and northeast.

"Bailey[1] explains this variation of snow on the two sides by saying that the greater part of the year the sun passes on the north side of the heavens from Arequipa, its rays striking with greater intensity on this side of the volcano."

Rondón was not satisfied with Bailey's explanation, and gave an explanation of his own, which seems to have a more reliable basis.

"We permit ourselves to give the following explanation for this phenomenon. During our stay of three days we have found frequently that with the temperature . . . standing at 8° to 12° Centi-

[1] H. D. Bailey was astronomer of the Harvard Observatory, and twice ascended the volcano.

grade, it suddenly went down to zero and below zero when the wind was blowing from the southwest side. And this sudden lowering of temperature occurred in spite of the fact that the sun was continuing to shine with all splendor. We believe that this constant current of air coming from the southwest brings the snows on this side."

An explanation similar to Bailey's has been used in other cases with little discrimination to account for the persistence of a greater depth of snow on the southern side of mountains in the southern hemisphere. In regions far to the south the solar influence may be the deciding factor, but in the latitude of Arequipa the prevailing direction of moisture-bearing winds is more important.

During my sojourn in Arequipa I observed that it was when the sun was almost at its farthest south, in early January, that snow began to gather on the southwestern side of the peak (the side facing Arequipa). This was when the rainy season was beginning. The sun passed the zenith at Arequipa on its return journey northward about February 4, but the rainy season continued until the early part of April, and snowfall on the peak likewise continued. The two phenomena came to an end practically at the same time.

During the height of the rainy season the usual daily sequence of weather around Arequipa was as follows: The early mornings were usually bright overhead. Clouds might conceal the tops of Chachani, El Misti, and Pichu Pichu from the first, but if they were not present they soon gathered. Winds or breezes blew from the south or southwest during the morning, and the clouds or overcast spread gradually from the mountains *toward* the direction from which the winds were blowing. Even after the sky overhead had become covered, the Cerros de la Caldera, along the southern horizon, and the desert region to the southwest in the neighborhood of Uchumayo and beyond were likely to be still in sunshine. Finally rain began to fall, beginning first in close proximity to the mountains. Sometimes it was in gentle showers and sometimes in a heavy downpour, accompanied by thunder and lightning. The rain might continue into the night or it might stop about sunset, and then the sun, sinking in the west below the clouds, illuminated their under surfaces and gave brilliant effects.

It does not seem difficult to explain these events. Moisture-laden air traveling inward from the ocean is forced upward by the rising land surface; it is cooled by adiabatic expansion and gradually reaches the saturation point. At Arequipa the inflowing air from the sea has attained a height of 7500 feet or more, and beyond that the rise along the flanks of the mountains is much steeper. Thus the southwestern side of the mountains receives the heavy precipitation. We have here an example of the well-known "rain-shadow" effect. Evidently, however, not all of the moisture is precipitated here, for the clouds, after passing over the dry northeastern slopes or escaping through gaps between the peaks, strike the main mass of the Cordillera, and more is brought down. Irregularities of the topography are likely to bring about local peculiarities.

Colton (1943, p. 641) has described and depicted an example of this cloud phenomenon on Mount Rumiñahui in Ecuador. Light (1941, p. 130), describing an airplane exploration of the Ruwenzori Range in Central Africa, speaks of clouds "forming out of clear air . . . that made them seem to advance with the speed of an express train."

On Mount Chachani deep glacial erosion has taken place on the southwestern side of the summit, but no information regarding conditions on the northeastern side has been found. On Pichu Pichu we have an excellent airplane photograph by Johnson (1930, p. 136). He states in the caption that,

"On Pichu-Pichu, as on El Misti, the heavier precipitation appears to be from the west. [It is actually the southwest that is shown in the photograph]. The eastern [northeastern] slopes have retained their conical form to a remarkable degree, while the west [southwest] side has been greatly modified by the action of ice and snow."

A crest of the mountain is shown, glaciated only on its southwestern side, and a glacial trough appears on this side.

The mechanism by which the southwestern sides of these volcanoes receive much more precipitation than the northeastern is understandable under the climatic conditions of the present time, but the astonishing thing is that a similar meteorological mechanism was operative during the glacial period and also during the long preceding pluvial period, during both of which eras the temperature and quantity of precipitation were far different from what they are at present.

In the Arequipa region, as in many others in both hemisphere where Pleistocene conditions have been studied, this period appears to have been characterized by increased precipitation as well as by lowered temperature. If, however, precipitation then was greater over certain large areas of the earth's surface than it is at present, a corollary seems to be implied that over other large areas evaporation was greater than normal in order to supply the increased precipitation, and hence in these latter areas the climate was warmer than normal. This seems at first sight an astonishing conclusion. The reasoning may be fallacious and should not be accepted without much investigation and consideration, but we may pursue the idea a little farther. Where were these warmer areas? Were temperatures, integrated over the whole earth's surface, actually so different from those of the present as has been assumed? We might propose the hypothesis that climatic conditions were far from steady in any one area, but were subject to large shifts, and that intervals of ameliorated conditions in some regions coincided with increased severity in others. The Pleistocene, then, may have been a period of sharper contrasts of climate and of shifting climates rather than a period of greater cold.

## REFERENCES CITED

Bowman, Isaiah (1916) *The Andes of Southern Peru*, Am. Geog. Soc., N. Y., 336 p.

Boyd, Louise A. (1935) *The Fiord Region of East Greenland*, Am. Geog. Soc., N. Y., 369 p.

Colton, F. B. (1943) Nat. Geog. Mag., vol. 84, no. 6.

Fenner, C. N. (1948) *Incandescent tuff flows in southern Peru*, Geol. Soc. Am., Bull., vol. 59, p. 879-894.

Johnson, G. R. (1930) *Peru from the air*, Am. Geog. Soc., 158 p.

Light, Richard V. (1941) *Focus on Africa*, Am. Geog. Soc., 228 p.

Ogilvie, Alan G. (1922) *Geography of the Central Andes*, Am. Geog. Soc., 240 p.

Rondón, M. Enrique (1937) *Expedición Universitaria del Año 1935 al crater del Volcán Misti*, in a collection of papers published by the University of Arequipa, Jan., 1937.

64 Broad St., Clifton, N. J.

Manuscript Received by the Secretary of the Society, July 12, 1946.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 919-926, 1 PL. SEPTEMBER 1948

# FURTHER OBSERVATIONS ON THE ROSS SHELF ICE, ANTARCTICA

BY ARTHUR DAVID HOWARD

## CONTENTS

| | Page |
|---|---|
| Abstract | 919 |
| Introduction and acknowledgments | 919 |
| Accumulation of snow and firn | 920 |
| Compaction | 920 |
| Differential motion | 921 |
| Density | 922 |
| Size and orientation of ice grains | 923 |
| Summary | 925 |
| References cited | 926 |

## ILLUSTRATIONS

| Plate | Facing page |
|---|---|
| 1.—Views in ice pit at Little America III, Bay of Whales, Antarctica | 922 |

## ABSTRACT

Observations on rate of accumulation, compaction, density, petrography, and differential motion of the firn in the upper levels of the Ross Shelf ice in the vicinity of Little America III were made during the Antarctic summer of 1946–1947. Studies were conducted in an enclosed pit dug to a depth of 23 feet by members of the expedition of 1939–1941. The bottom of this pit is now about 36 feet below the surface. The present observations project our knowledge of the physical characedge of the physical characteristics of the Ross Shelf ice to a greater depth than heretofore investigated.

Eighteen feet of firn accumulated during the period 1940–1947. The annual rate of 31 inches thus indicated must have been greater originally as subsequent compaction has taken place. The compaction was greatest in the upper, more porous layers. Petrographic examination indicates that the density increases were primarily the result of this compaction. No evidence of recrystallization was observed, except for occasional interpenetration of grains at the deepest level. This would seem to restrict recrystallization to depths greater than 36 feet. There was no evidence of differential motion of the firn to this depth.

## INTRODUCTION AND ACKNOWLEDGMENTS

During the U.S. Navy Antarctic Expedition of 1946–1947 limited observations on the physical character of the Ross Shelf ice were made. Attention was confined to an enclosed ice[1] pit, 2 meters square and 7 meters deep, which had been dug by mem-

[1] Actually, all of the Ross Shelf ice above sea level in the area studied consists of firn (névé). Firn is generally considered to differ from true ice in the large number of connected air spaces and a density less than 0.80. The term ice is used in the present report as a matter of convenience. Unless otherwise stated, firn is implied.

bers of the expedition of 1939–1941. At that time, observations had been made of density, rate of settling, differential movement, size and orientation of ice crystals, and stratification (Wade, 1945). Observations on temperature and rate of accumulation of show had been made outside the pit.

The observations herein recorded were made under severe time limitations, hence are not as detailed as would otherwise have been possible. They do indicate, however, the types of changes that occur in the upper levels of the shelf ice over a period of years, and project our knowledge of the physical characteristics of the ice to a deeper level.

The writer is indebted to Dr. Paul Siple, Mr. James Balsley, and Mr. Amory Waite, who graciously took time off from their own studies to assist in some of the measurements.

## ACCUMULATION OF SNOW AND FIRN

In January, 1947, all of the buildings of the expedition of 1939–1941 (Little America III) were completely buried, only masts and ventilators projecting above the snow. Entry was made through a hatch, 3 to 4 feet below the surface.

Measurements indicate that 18 feet of snow and firn had accumulated in the 7 years since the buildings were constructed. This indicates an annual rate of accumulation of 31 inches. However, the rate must have been somewhat greater inasmuch as compaction undoubtedly took place. On the other hand, the buildings may have caused a greater local accumulation of drift, although a swell in the topography could not be detected. In any event, the rate of accumulation for the period 1940–47 checks closely with the findings of the previous expedition, when observations made at several different stations showed an average residual increase of 84 cm. (33 inches) over a period of 348 days. Actually, the total accumulation had been twice as great, but half had been removed by deflation and other wasting processes.

The average of 2½ to 3 feet per year cannot be applied indiscriminately to the entire area of the Ross Shelf or to any one spot throughout time. Byrd reported (1935, p. 72) that in January 1934 the ground level on which the buildings of Little America I had been built in January 1930 was buried under 14 feet of snow. Presumably the term snow was meant to include firn. This would indicate an average annual accumulation, after compaction of 3½ feet. This site is in a shallow valley. Little America II was built in 1934 on the surface above Little America I. A rough estimate of the annual rate of accumulation since that time is 1½ to 2 feet. In the McMurdo Sound area, across the Ross Sea, an annual increase of only 15 inches over a period of 6½ years was noted (Wright and Priestley, 1922, p. 162).

## COMPACTION

The ice (firn) pit (Pl. 1, fig. 1) had originally been dug to a depth of 7 meters (22.9 feet). Remeasurement shows it now to be only 5 meters (16.4 feet). However, there is an accumulation of about 2 feet of ice rubble in the base of the pit. Hence its present depth is about 18½ feet. An independent estimate of the amount of compaction has been made by re-examination of the compression meters left behind from

the last expedition. The three compression meters (Pl. 1, fig. 2), of a type modified by Moss (1938, p. 219–220) from an earlier model, were installed in the north wall of the shaft and measured the settling in overlapping zones, from −0.5 to −2.5 meters (1.6 to 8.2 feet), from −1.5 to −3.5 meters (4.9 to 11.5 feet), and from −2.5 to −4.5 meters (8.2 to 14.8 feet). Thus, each compression meter measured the compaction in a zone 2 meters (6.6 feet) thick. The uppermost instrument was in place the longest, 190 days, giving complete records from July 1940 to January 1941. All three instruments were found broken when examined in January 1947. However, remeasurement of the vertical intervals between their basal blocks probably gives a fairly accurate measure of the total compaction within the range of each instrument. Remeasurement of the original 0.5 meter interval between the rim of the pit and the top of the first compression meter was not attempted because of the large error which would have been introduced in measuring so small an interval from the irregular rim of the pit.

The three 2-meter intervals were found to have diminished to 1.61, 1.63, and 1.65 meters respectively, from top to bottom. Thus, within the span of the upper instrument there has been 0.39 meters (19.5%) compaction, within the span of the middle instrument there has been 0.37 meters (18.5%) compaction, and within the span of the lower instrument, 0.35 meters (17.5%) compaction. The greatest amount of compaction took place in the upper levels, as was to be expected.

Within the walls of the pit are a series of cylindrical holes from which samples had been taken on the last expedition. These holes are now elliptical in cross-section, those near the rim of the pit showing two-thirds flattening, those near the bottom showing one-half flattening. This checks with the data furnished by the compression meters relative to greater rate of compaction at the top of the pit. The amount of flattening of the holes, however, does not furnish absolute data on the amount of compaction. Closing of the upper holes to one-third of their original height, and the lower to one-half their original height, would mean a contraction of the depth of the entire pit of possibly 55 to 60% i.e., 12.6 to 13.7 feet. Yet measurement shows that the maximum reduction in depth of the pit has been only about 4 to 4½ feet, or 17 to 20%. Hence, the holes must close by local settling.

## DIFFERENTIAL MOTION

Surveys conducted during the three previous Byrd expeditions indicate that the shelf ice in the vicinity of Little America is moving westward. To determine whether there was any differential motion in a vertical plane, Wade (1945, p. 166) inserted vertical rows of closely spaced pins in the walls of the shaft. No differential movement was indicated over the 3-month period during which the observations were made. Unfortunately, these vertically arranged pins were not found in January 1947. Hence, the only information bearing on the problem of differential motion is indirect.

A *horizontal* row of closely spaced pins was found in the east wall. These now descend slightly to the north. In the west wall, the long axis of the elliptical cross-sections of some of the higher sample holes mentioned earlier dip to the north. The

rim of the shaft itself is about 1 foot lower in the northwest corner than in the southeast corner. All these observations are interpreted to mean only that the northwest corner of the building housing the ice pit has settled most under the weight of the later snows. The walls of the shaft, on the other hand, although rough, show little if any deviation from the plumb line. It is therefore assumed that the shelf ice has moved *en masse* during the last 7 years, at least in its upper levels. This agrees with Wade's (1945, p. 166) original conclusion, but projects the data down to a depth of over 36 feet.

For the assistance of geologists of future expeditions, two vertical rows of 6-inch nails have been driven in the north and east walls of the shaft (Pl. 1). They were oriented by plumb line and spaced by tape. The line of spikes in the east wall is about 2 feet out from the northeast corner. The first spike was driven 1 foot below the base board at the rim of the pit, and subsequent spikes are 1 foot apart. Due to the irregularity of the wall, there may be an error of about $\frac{1}{4}$ inch between each pair of spikes, and a similar deviation from the plumb line. In no case is an error of as much as $\frac{1}{2}$ inch probable. Vertical intervals were measured from the top down to each new spike, not from spike to spike. Hence, errors in vertical interval are not cumulative. The line of spikes in the north wall, about 6 inches out from the northeast corner, are similarly emplaced, except that because the rim of the pit descends toward the northwest corner, the top spike has been driven in at about the same level as the top spike in the east wall. Thus, the spikes in the two walls line up horizontally fairly closely. Each line includes 15 spikes.

## DENSITY

Four specimens were taken from the east wall of the pit at 5-foot intervals, starting 6 inches below the rim and terminating 1 foot above the ice rubble at the base. Since these specimens had to be stored in airtight containers for return to the ship where density determinations could be made, it was not possible to collect specimens as large as desired. Specimen No. 4, taken from the highest level, proved to be firmest and was sawed and sand-papered with great care to a $1\frac{1}{2}$ inch cube. Of the others, Specimen No. 3, the next lowest, was not quite as firm, and Specimens No. 2 and No. 1 were somewhat brittle. Specimens 1 to 3 had to be reduced to 1 x $1\frac{1}{2}$-inch blocks before sharp forms could be secured. Each sample was a trifle under volume because of faint rounding of the corners. The percentage loss in volume was estimated for each sample and the density recalculated. Thus, a range of density is given for each specimen. The derived densities are tabulated below.

DENSITY OF FIRN SPECIMENS FROM PIT AT LITTLE AMERICA III

| SPECIMEN NUMBER | DEPTH BELOW RIM OF PIT IN FEET | DEPTH BELOW PRESENT SNOW SURFACE IN FEET | DENSITY (SPECIFIC GRAVITY) |
|---|---|---|---|
| 4 | 0.5 | 18.5 | 0.404–0.416 |
| 3 | 5.5 | 23.5 | 0.419–0.432 |
| 2 | 10.5 | 28.5 | 0.542–0.548 |
| 1 | 15.5 | 33.5 | 0.531–0.542 |

FIGURE 1. ICE PIT AT LITTLE AMERICA III

Sides of pit are still essentially plumb, indicating little or no differential movement of the ice. Note broken compression meters in north wall (top of photo). Upper end of ladder became wedged under a winch at top of pit and broke as pit contracted by compaction. Vertical row of spikes in east wall were emplaced by the writer for future observers. A second row of spikes is partly hidden by the ladder. Official U. S. Navy photograph.

FIGURE 2. BROKEN COMPRESSION METERS IN NORTH WALL OF ICE PIT AT LITTLE AMERICA III

As designed, vertical rods between wooden blocks are slightly offset, so that their ends move past each other as the distance between the blocks diminishes. Amount of movement is recorded by horizontal lever moving across vertical scale built out from lower rod. Official U. S. Navy photograph.

VIEWS IN ICE PIT AT LITTLE AMERICA III, BAY OF WHALES, ANTARCTICA

Density observations made during the previous expedition also showed a general increase downward, but with many variations from a smooth-line curve. Actually, the lowest density recorded at that time was not near the surface (0.360 at 0.2 meters), but several meters below (0.325 at 2.9 meters). Eleven out of twelve samples down to a depth of 2.0 meters showed densities of less than 0.4, seven out of ten samples between 2.0 and 3.65 meters had densities between 0.4 and 0.5, while eight out of nine specimens below 3.65 meters showed densities greater than 0.5. The maximum density recorded was 0.569 at 5.1 meters, although measurements were made down to 5.8 meters.

The four samples taken from the pit during the present expedition show about what was to be expected, that the upper layers have suffered a greater increase in density than the lower. Specimen No. 3, and probably also Specimen No. 4, were taken from stratigraphic levels which, 6 years ago, had densities less than 0.4. Now the densities are greater than 0.4. The densities of the two lower specimens, however, are within the range of the original variations for those depths.

It has been noted that 18 feet of snow and firn had accumulated during the period 1940–1947. Probably not much more than half of this is pressing down on the building which houses the ice pit. Hence, the ice on which the building stands has not been subjected to the full weight of the 18 feet of overburden. Both the compaction and density measurements, therefore, are undoubtedly lower than elsewhere. Dr. Paul Siple and Major James Holcombe determined the density near the base of the barrier where the barrier was quite low. The specimen was taken from about 2 feet above the water line and at an estimated distance of 35 feet below the rim of the barrier. The density was determined to be 0.64, a minimum, rather than an average or maximum figure.

Poulter (1947, p. 170) reported that the average elevation of the surface of the shelf ice around the Bay of Whales is 94 feet and its average thickness about 760 feet. This would indicate an average density of about 0.84 for the shelf ice in this area. It is interesting to note that Perutz and Seligman (1939, p. 344) found a density of 0.80 at a depth of 23 meters (75 feet) in the Great Aletsch Glacier. At this depth the amount of air space had decreased appreciably and the spaces were well separated from one another. Hence, Perutz and Seligman called this specimen true glacier ice. Although the high density at this depth in the Great Aletsch Glacier may be due to differential internal movements, the rate of density increase in the Ross Shelf Ice suggests that its average density of 0.84 may be reached within a relatively short distance of the surface, perhaps 100 feet, by simpe compaction.

## SIZE AND ORIENTATION OF ICE GRAINS

Investigation of size and orientation of ice grains with depth was confined, during the 1939–41 expedition, to the uppermost 2 meters of the firn, and to thin bands of ice down to 5 meters. Warner's (1945, p. 173) conclusions were as follows: "There is a general trend toward increase in grain size with depth but with notable exceptions. It is safe to say that there is no definite orientation of ice crystals, at least to the depth investigated." Wade (1945, p. 166) pointed out that Perutz and Seligman (1939, p. 341–345) did find a definite crystallographic orientation of grains at shallow

depths in the firn of the Great Aletsch Glacier. According to these authors, such orientation results from melting and refreezing or sublimation. To account for the absence of orientation in the firn of the Ross Shelf Ice, Wade suggested that meltwater is the essential requisite and remarks that no meltwater is present in the firn of the shelf ice during the entire year. It is by no means established, however, that meltwater is always absent. Weather observations made by personnel of the Byrd expeditions have been both careful and complete, but, gaps of several years separate each series of observations. Temperature conditions in these intervals are unknown. Furthermore, the records of two of the three previous expeditions do show temperatures exceeding 32°F. Unfortunately maximum temperatures were not recorded in the summer of 1929–1930.

Grimminger and Haines (1939, p. 100–103) recorded 3 days in December 1934 and 6 days in January 1935 when the maximum temperature exceeded 32°F. The high was 37.9° on December 17, 1934. On one of these days there was light rain, and on another, moist snow. Court (1945, p. 332), meteorologist with the expedition of 1939–1941, reported that there were 4 days in late December 1940 and 1 day in early January 1941 in which the daily *mean* of the maximum and minimum temperatures exceeded 32°F. He also states (verbal communication) that there were 7 days in December and 3 days in January when the maximum temperature exceeded the melting point. The highest temperature recorded was 42.7°F. Observations were taken every 6 hours. Altogether, 18 observations exceeded 32°, 13 in December and 5 in January. Court reports that all these maxima occurred with overcast conditions, and nearly all with the sun at upper transit. Furthermore, the wind was usually from the north, that is, off the sea, and 3 of the maxima were recorded while a mixture of rain and snow was falling. Court believes, however, that the maximum readings are probably too high because the instrument shelter absorbed heat. He doubts that the true air temperature was ever more than a few degrees above freezing and then within only a relatively few miles of the coast.

The fact that the temperature does reach the melting point, however, suggests that meltwater may sometimes be present in the firn of the shelf ice. Court did not observe any during his stay in the Bay of Whales area. Nor was any observed by the writer during the present expedition, although the continent was not reached until after the peak of the summer. It may be, however, that the meltwater consists of nothing more than films of molecular thickness, which do not saturate the firn to the extent that it feels wet.

In any event, until it is certain that meltwater never forms in the firn of the Ross Shelf ice, it seems best not to employ either its presence or absence as an argument in connection with the problem of crystallographic orientation. Other factors that may be involved are the rapidity of temperature changes and the slope of the temperature gradient.

The writer undertook further petrographic examination in this area because the ice pit in which the original studies were made in 1940–1941 was now (1947) 18 feet farther below the surface. It was felt that melting and refreezing, or sublimation, might take place in depth due to the weight of the overlying snow and might cause a crystallographic orientation, even though one or both of these processes might not

operate at the surface. It should be noted however, that Perutz and Seligman (1939, p. 344–345) found that in the Great Aletsch Glacier crystallographic orientation disappeared with depth due to small scale, internal adjustments.

Two thin-sections, normal to each other, were prepared from each of the four specimens taken from the pit by the writer. The sections were prepared as follows: 1) The specimens, because of their relative friability, were immersed in water at about the freezing point for a few moments and then allowed to freeze solid. 2) A 1½-inch oriented cube was cut, making use of a double-bladed hack saw devised by the writer, in which the blades were set 1½ inches apart. Cubes of this size were also used in storage and density tests. 3) The cube was set in a 1½-inch square recess in a board, the recess being ¼ inch deep. The lower ¼ inch of the cube was then sawed off. Perforations in the base of the recess permitted the specimen to be prodded loose. 4) One side of the ¼-inch thick section was then polished flat, using three different grades of emery paper, and cemented with water to a 2 x 3-inch glass slide. 5) The upper surface of the specimen was then shavd with a sharp knife to a thickness of about 1 to 2 mm., and sandpapered to about 0.5 mm. No cover glasses were used. The entire operation took about 30 minutes. The sections were then examined, first by means of non-magnifying polaroid plates, and then under the petrographic microscope.

All four specimens were remarkably alike. Grain size averaged about 1 to 1½ mm., except for Specimen No. 3 which averaged 1½ to 2 mm. Rarely were grains up to 3 mm. observed. The granules were equidimensional, almost without exception. The fabric was allotriomorphic, although intergranular contacts tended to be subangular. In spite of the latter characteristic, the resemblence to a colorless, perlitic glass was remarkable. Grain boundaries were frequently outlined by elongate air bubbles or by strings of tiny bubbles. Some of the elongate bubbles exceeded 1 mm. in length. In general, however, the granules were jammed tightly together, and occasional grains dovetailed with one another in such a way as to suggest mutual penetration. This may indicate the beginning of pressure effects whereby solution or evaporation occurs at the points of greatest pressure with recrystallization elsewhere. Specimen No. 3 showed one excellent example of irregular interpenetration of two grains. Very few grains were elongate, and nothing resembling schistosity was observed.

The microscopic examination thus seems to indicate that the pressure resulting from the weight of the overlying firn has had no effects on even the deepest of the samples other than increased compaction, flattening and occasional interpenetration of grain boundaries, and elongation and comminution of air bubbles. No elongation of grains normal to the downward directed stress was observed. If such occurs in the ice of the Ross Shelf, it is at depths below 36 feet.

## SUMMARY

Under the weight of 18 feet of snow and firn, representing the accumulation over the period 1941–1947, 23 feet of firn was compacted to approximately 18½ feet. As a result, further increases in density have occurred, with the upper layers showing the greatest increase. It was found, 6 years ago, that the density of the firn does not

increase uniformly downward, but shows numerous anomalies. Inasmuch as the layers from which density determinations were made on the present expedition could not be correlated exactly with the layers examined 6 years ago, the density determinations are indicative only of trends. Suffice it to say that specimens taken from the stratigraphic zone in which, 6 years ago, densities varied between 0.3 and 0.4, now give densities exceeding 0.4. Densities of the lower specimens fall within the range observed 6 years ago, but a measurement made outside the ice pit gave a density of about 0.64 at approximately 35 feet below the surface. The average density of the Ross Ice Shelf in the vicinity of the Bay of Whales is probably about 0.84, judging from data on thickness and elevation of the Shelf furnished by Poulter (1947).

No differential motion of the shelf ice was noted, other than that due to greater local settling of one corner of the ice pit. Thus, the conclusions reached by Wade in 1941, that so far as his observations to a depth of 23 feet were concerned, the Ross Shelf ice was moving forward as a unit, is here confirmed and extended to a depth of 36½ feet.

No elongation or flattening of grains due to the pressure of accumulating snows was noted. Wade's conclusion that the increase in density down to a depth of 23 feet was due to mechanical settling and compaction may, therefore, be extended to 36½ feet. However, several examples of intergranular penetration suggest that solution and/or evaporation along grain boundaries may be commencing at this depth.

## REFERENCES CITED

Byrd, Richard E. (1935) *Discovery*, G. P. Putnam's Sons, New York.

Court, A. (1945) *Weather Observations during 1940–1941 at Little America III, Antarctica*, Reports on Scientific Results of the U. S. Antarctic Service Expedition, 1939–1941, Am. Philos. Soc., Pr., vol. 89, p. 160–173.

Grimminger, G. and Haines, W. C. (1939) *Meteorological Results of the Byrd Antarctic Expeditions, 1928–30, 1933–35*, Tables, Monthly Weather Review, Supplement No. 41, U. S. Weather Bureau, Washington.

Moss, R. (1938) *The Physics of an Ice Cap*, Geog. Jour., vol. 92, p. 211–227.

Perutz, M. F. and Seligman, G. (1939) *A Crystallographic Investigation of Glacier Structure and the Mechanism of Glacier Flow*, Royal Soc. London, Pr., vol. 172, p. 335–360.

Poulter, T. C. (1947) *Seismic Measurements on the Ross Shelf Ice-Part* I, Am. Geophys. Union, Tr., vol. 28, no. 2, p. 162–170.

Wade, F. A. (1945) *The Physical Aspects of the Ross Shelf Ice*, Reports on Scientific Results of the U. S. Antarctic Service Expedition, 1939–1941, Am. Philos. Soc., Pr., vol. 89, p. 160–173.

Warner, L. A. (1945) *Summary of Crystallographic Investigations of the Ross Shelf Ice.* Appendix in Wade (1945), *The Physical Aspects of the Ross Shelf Ice*, Am. Philos. Soc., Pr., vol. 89, p. 172–173.

Wright, C. S. and Priestley, R. E. (1922) *British (Terra Nova) Antarctic Expedition, 1910–1913*, Glaciology, London.

Stanford University, Stanford, California.

Manuscript Received by the Secretary of the Society, July 11, 1947.
Published by Permission of the Director, U. S. Geological Survey.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 927-940, 3 FIGS., 1 PL. SEPTEMBER 1948

# HIGH-POTASH DIKES IN THE HOMESTAKE MINE, LEAD, SOUTH DAKOTA

BY JAMES A. NOBLE

## CONTENTS

| | Page |
|---|---|
| Abstract | 927 |
| Introduction | 927 |
| Tertiary igneous rocks of the northern Black Hills | 928 |
| Rhyolite dikes in the Homestake Mine | 930 |
| Other occurrences of high potash dikes | 934 |
| Origin of the high potash rhyolite dikes in the Homestake Mine | 935 |
| References cited | 938 |

## ILLUSTRATIONS

| Figure | Page |
|---|---|
| 1. Ternary diagrams of normative mineral compositions of rhyolite dikes | 933 |
| 2. Intersections of rhyolite dikes | 934 |
| 3. Ternary diagram of the system Or-An-Ab | 936 |

| Plate | Facing page |
|---|---|
| 1. Photomicrographs of typical sections of rhyolite from the Homestake mine | 930 |

## ABSTRACT

Widely scattered bodies of Tertiary intrusive porphyry crop out in the northern Black Hills, in greatest abundance in a circular area of about 10-mile radius with Lead and the Homestake Mine at the center. Five rock types are represented: rhyolite; quartz monzonite; monzonite; grorudite; and phonolite. A swarm of rhyolite dikes in the Homestake Mine comprise rocks of two distinct compositional types: so-called low potash rocks with about equal parts of orthoclase, albite, and quartz; and high potash rocks which consist of orthoclase and quartz. There are no ferro-magnesian minerals in either type of rock.

While only a few other occurrences of rocks of the high potash type have been described, it does not seem possible to dismiss this rock type from a genetic classification on the basis of inaccurate description. Three possible modes of origin for similar rocks have been proposed: selective assimilation of potash-rich wall rocks; hydrothermal introduction of potash; or, most probably, normal differentiation with strong reaction, followed by hindered reaction (Bowen, 1928).

## INTRODUCTION

This paper discusses the petrology of some Tertiary intrusive rocks in the northern Black Hills, particularly some high potash dikes which are found in the Homestake Mine and its immediate vicinity near Lead, South Dakota. Only the petrology is considered; the structural relations of the intrusive rocks to each other and to the

intruded rocks will be discussed separately. This manuscript has been read critically by Prof. E. S. Larsen of Harvard University, Dr. D. H. McLaughlin, president of the Homestake Mining Company, Dr. R. G. Wayland, and Mr. J. O. Harder, to all of whom I am indebted for assistance. For opportunity to carry on the work and for defraying costs of analyses, I am indebted to the Homestake Mining Company.

The Tertiary igneous rocks of the northern Black Hills have been described by several writers. The descriptions by N. H. Winchell (Ludlow, 1875, p. 21–66) derived from the rapid reconnaissance of the Custer expedition were necessarily very meager. The field descriptions by Newton and Jenney (1880, p. 189–203) were not much more complete, but J. H. Caswell's (Newton and Jenney, 1880, p. 481–527) descriptions of their specimens were sufficiently accurate and detailed to subdivide the rocks into the main groups. Irving (1899) made a much more complete classification after studying a part of the northern Black Hills. A study by Jaggar (1901) emphasized the structure and mode of origin of the intrusive bodies but discussed the petrography only briefly. Paige (Darton and Paige, 1925, p. 16–17) classified the Tertiary igneous rocks and gave brief but accurate descriptions of each group.

## TERTIARY IGNEOUS ROCKS OF THE NORTHERN BLACK HILLS

The Tertiary igneous rocks form stocks, dikes, sills, possibly laccoliths, and two or three small bodies of extrusive lava. They are exposed only in the northern portion of the Black Hills. Custer Peak, about 8 miles south of Lead, and Inyankara Mountain, 32 miles southwest of Lead, are the most southerly bodies exposed. The most easterly occurrence is Bear Butte, 20 miles northeast of Lead, and the most westerly one is Missouri Buttes, 50 miles northwest of Lead. The greatest abundance of Tertiary igneous rocks is in a circular area of about 10-mile radius with Lead at the center. The more distant igneous masses west of the South Dakota boundary have not been studied.

The Tertiary igneous rocks comprise five separate but closely related types. In probable order of abundance, beginning with the most abundant, these are rhyolite, quartz monzonite, monzonite, grorudite, and phonolite. Most are porphyritic. There seem to be gradations from one type to another.

The rhyolites, all light in color, can be subdivided into three groups on the basis of the phenocrysts. One group has only feldspar phenocrysts, large and abundant in a few, but in most small and not very abundant, and in some procelainlike rocks, scarcely visible. A second group has phenocrysts of quartz, either in small, rounded, partly resorbed grains, or, in a few instances, in moderately large bipyramids. These quartz porphyries have a few small feldspar phenocrysts, and by increase of the amount of feldspar they grade to the third group, quartz-feldspar porphyries, in which the phenocrysts are quartz and feldspar in about equal amounts. Biotite is a minor constituent of most of the rhyolites, except those in the Homestake Mine, and accessory garnet is visible in some rocks. The quartz porphyries form small stocks and large dikes and sills in the vicinity of Lead and elsewhere in the northern

Hills, but the feldspar porphyries are the most abundant and widely distributed rocks.

Under the microscope the feldspar phenocrysts of the rhyolites are found to be orthoclase, microcline, albite, or two or three of these feldspars, with a few grains of oligoclase in some specimens. A few nearly completely resorbed quartz phenocrysts are seen in some rocks in which they are not visible in hand specimen. The quartz phenocrysts of most of the quartz porphyries are partly resorbed by the groundmass. Biotite is moderately abundant in some rhyolites, and in some of these it shows a poor orientation into a platy flow structure. The groundmass of the rhyolites is a fine-grained to cryptocrystalline aggregate of orthoclase, quartz, and biotite. There is a little accessory sphene, apatite, and magnetite. Some rhyolites have a trachytic texture.

The quartz monzonites are on the whole somewhat darker than the rhyolites but cannot be distinguished surely on the basis of color. They have abundant phenocrysts of feldspar and a few have quartz in addition. Dark minerals are biotite and hornblende. Under the microscope, phenocrysts are seen to be orthoclase, oligoclase, andesine, quartz, and hornblende. Minerals of the groundmass are orthoclase and plagioclase feldspars, quartz, biotite, and accessory magnetite, sphene, apatite, and zircon. Quartz monzonites are not abundant, but some may have been overlooked, as they are not readily distinguished from rhyolites and from grorudites.

The monzonites are gray, seriate to epigranular, and differ from the quartz monzonites in the absence of quartz phenocrysts and the virtual absence of quartz from the groundmass. Hornblende is abundant in some monzonites. The distribution of these rocks is not wide, but one stock near Lead is monzonite.

Grorudites form many dikes and sills in the vicinity of Terry Peak and a few in other parts of the northern Hills. They can be distinguished from the rhyolites by their darker color. Some varieties are green, but gray varieties are not readily distinguished in the field from monzonites or phonolites. Many of the grorudites give off a strong fetid odor when freshly broken or when struck a sharp blow. Irving (1899, p. 278) has noted the same peculiarity in a rock which he classed as rhyolite. The grorudites differ chemically from the rhyolites in having a higher soda content and mineralogically in the predominance of albite and decrease in quartz. Phenocrysts are albite and some microcline, orthoclase, quartz, and aegirite or aegirite-augite. In some varieties the phenocrysts are very sparse but in others they are abundant. The groundmass of most of the grorudites is composed of albite and a little orthoclase, apatite, magnetite, biotite, and quartz, but some have no quartz and are better described as albite syenites. Paige (Darton and Paige, 1925, p. 16) has subdivided the grorudites on the basis of the arrangement of the aegirite phenocrysts.

The phonolites are dark-brownish gray, greenish gray, and green rocks with fairly conspicuous phenocrysts. They are almost wholly confined to the central portion of the intrusive area, between Englewood and Spearfish Peak. Phenocrysts comprise orthoclase and some albite, aegirite-augite, aegirite, and biotite. In the groundmass are orthoclase, albite, aegirite, and, in one specimen or another, sodalite,

analcite, nepheline, nosean, sphene, and apatite. Zeolites replace the feldspars in some rocks.

Both Jaggar (1901, p. 183–185) and Paige (Darton and Paige, 1925, p. 16) note inconsistencies in the age relations of the Tertiary igneous rocks, and the age relations need more study. In many instances the phonolites are the latest rocks, but some phonolites are cut by rhyolite. Near Roubaix, about 7 miles southeast of Lead, is a small area of rhyolite lava and breccia, probably of Oligocene age, as first described by Darton (1912), not Quaternary, as described later (Darton and Paige, 1925, p. 17). It probably rests on and is contemporaneous with Oligocene gravel deposits and an Oligocene erosion surface. This rhyolite lava and breccia is the youngest of the Tertiary igneous rocks.

The petrology of the Tertiary igneous rocks of the northern Black Hills has not been adequately studied. The origin and true relations of the several types of rocks will not be fully known until the igneous bodies are mapped on a scale large enough to show each separate body and each one has been studied microscopically.

## RHYOLITE DIKES IN THE HOMESTAKE MINE

Only three members of the group of Tertiary igneous rocks of the northern Black Hills are found in the Homestake Mine: namely, rhyolite; phonolite; and grorudite. The rhyolite forms a well-defined zone of dikes and a few individual dikes outside that zone, and it also forms thick sills which cap the hilltops on each side of the Homestake open cut. The phonolite forms one dike in the mine and two or three small dikes on the surface near the open cut. In an outlying part of the mine are two narrow dikes of grorudite.

All the rhyolite dikes which are found in and near the Homestake Mine strike a few degrees west of north. They strike a little nearer to north than the average strike of the schists and are much straighter than are the schists, although locally the dikes are parallel to the schist bedding. Although a few dikes dip west, most of them dip east at angles ranging from 60° to nearly vertical. In a few places it was determined that quartz porphyries are later than feldspar porphyries. On the other hand, the dikes of feldspar porphyry on continuing upward above the base of the Cambrian strata form sills which definitely cut the thick sills of quartz porphyry on the sides of the open cut. The phonolite dike in the mine and those on the surface near the open cut have a totally different attitude. They strike about 50° to 60° east of north and dip north at about 35°. The phonolite dike in the mine is later than the rhyolite dikes.

The rhyolite dikes in the Homestake Mine are white, buff, and gray. Some hard, dense, porcelainlike rocks break with a conchoidal fracture. All the rhyolites have feldspar phenocrysts, and some have quartz phenocrysts, but in some the phenocrysts are almost absent. No ferro-magnesian minerals can be seen in the hand specimen. Most dikes are well banded for an inch or two along the walls. Where the walls of the dikes are composed of a series of scallops between schist cusps, the banding narrows to an inch or less over the schist cusps and widens to as much as 6 inches in the scallops. Blunt-nosed dike endings and dike fingers are commonly,

FIGURE 1. MICROGRANULAR TEXTURE IN RHYOLITE

Consists of orthoclase, quartz, and very little pyrite and calcite. Minerals are in small anhedral grains. In this specimen orthoclase is estimated to form 80–90 per cent of the rock. Plane polarized light.

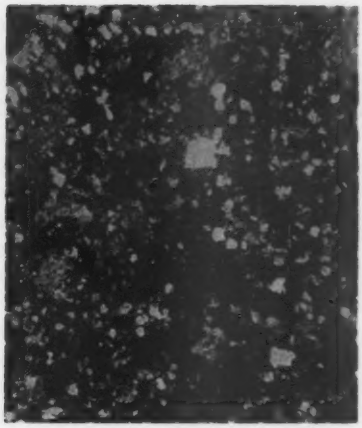


FIGURE 2. SAME FIELD AS 1, BUT X-NICOLS.

FIGURE 3. MICROPOIKILITIC TEXTURE IN RHYOLITE

Tiny laths of orthoclase enclosed in large very irregular grains of quartz. Large grains of quartz make complete background for small laths of orthoclase. Very little pyrite and calcite. By analysis the rock contains 11.15% $K_2O$ and by calculation 68.5% orthoclase, 4.6% albite, and 25.3% quartz (*see* analysis 6). Plane polarized light.

FIGURE 4. SAME FIELD AS 3, BUT X-NICOLS.

PHOTOMICROGRAPHS OF TYPICAL SECTIONS OF RHYOLITE FROM THE HOMESTAKE MINE
ALL X 150

t not invariably, banded for widths of several inches or even a few feet. Small oliths of schist are surrounded by narrow banded zones.

The microscopic descriptions are based on a study of more than 100 thin sections. the rhyolites are characterized by an extremely fine-grained crystalline to crypto-stalline groundmass. In most dikes there are only four or five phenocrysts in ordinary thin section, and no phenocrysts exceed 2.5 mm. long. Orthoclase is e most common phenocryst, but there are a few of albite and a very few of oligo-se-andesine. In a few dikes there are small, partly resorbed quartz phenocrysts. he dikes are composed wholly of feldspars and quartz, and the microscope shows t a single dark mineral. Two distinct textures are microgranular and micro-ikilitic (Pl. 1). In the microgranular texture, subhedral to anhedral grains of thoclase, albite, and quartz, 0.1 mm. or less in diameter, form a fine-grained mosaic. the micropoikilitic texture, anhedral or nearly round grains of quartz, orthoclase, albite about 0.3 mm. in diameter enclose tiny unoriented laths of orthoclase and rhaps albite from 0.01 to 0.1 mm. long. In those instances where relative ages uld be determined, dikes with micropoikilitic texture were later than dikes with icrogranular texture. It has not been found possible to correlate the textures with y physical properties in the hand specimens.

No dikes are wholly free from hydrothermal alteration, but the amount of altera-on is not great except in a few places. Pyrite is widely distributed in tiny cubes nd groups of grains, and calcite replaces the phenocrysts and the groundmass. A ttle sericite replaces the groundmass of some dikes but is not found in most of hem.

The marginal banding is scarcely visible under the microscope. It seems to be ue to alternating bands of coarse and fine grain, to small amounts of inclusions erived from the schists, and to selective replacement by calcite. There may be a oncentration of fine-grained orthoclase at the very contact, but the mineral iden-ification is not sure because of the exceedingly fine grain size.

In the schist wall rocks there are many veinlets of calcite and pyrite, but in most laces there is little or no thermal metamorphism in the schists. Some specimens hich actually adjoin rhyolite show no alteration even where the dike contact is isible in the thin section. A very few sections from points within a few inches of he contact show small blebs of very fine grained orthoclase enclosed in the schist minerals. But hydrothermal minerals, pyrite, sphene, magnetite, specularite, alcite, quartz, and sericite, replace the schists. These minerals, and some galena, phalerite, chalcopyrite, and fluorite, also form tiny veins in the schists and in the ikes.

The rhyolite dikes of the Homestake Mine are characterized by abnormally high potash content. On the basis of chemical composition the rocks have been separated into two types, called for convenience high-potash and low-potash types, but it will be noted that even the low-potash type has an unusually high potash-soda ratio. There are no rocks of intermediate character. All the available analyses (except some of composite samples made by Sharwood (1911)) belong to one or the other of these two distinct groups (Table 1; Fig. 1). Six of the analyses listed in Table 1 and Figure 1 are specimens of rhyolite collected just at the dike wall. These are

plotted by crosses, and on the diagram of Figure 1-C they are connected to the analyses of specimens from the central portion of the same dike by dashed lines. The material selected for analysis, except the specimens used by Sharwood, was

TABLE 1.—*Analyses of rhyolite dikes of the Homestake Mine*

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **HIGH-POTASH RHYOLITES** | | | | | | | | | | | | | | |
| $SiO_2$ | 70.26 | 68.10 | 68.54 | 68.82 | 70.72 | 70.10 | 70.96 | 70.04 | 71.90 | 69.20 | 69.96 | 57.74 | 61.54 | 67.08 |
| $Al_2O_3$ | 14.38 | 14.42 | 14.13 | 14.08 | 15.18 | 14.51 | | | | | | | | |
| $Fe_2O_3$ | tr | 0.80 | 2.63* | 2.54* | | | | | | | | | | |
| FeO | | | | | | 0.77 | | | | | | | | |
| MnO | 0.06 | 0.06 | | | | | | | | | | | | |
| MgO | 0.45 | | 0.22 | 0.39 | | 0.82 | | | | | | | | |
| CaO | 0.63 | 0.75 | 0.68 | 0.88 | 1.24 | 1.18 | | | | | | | | |
| $Na_2O$ | 0.40 | 0.28 | 0.32 | 0.32 | 0.59 | 0.53 | 0.35 | 0.23 | 0.31 | 0.24 | 0.36 | 0.39 | 0.47 | 0.34 |
| $K_2O$ | 12.12 | 12.13 | 11.11 | 10.04 | 9.17 | 11.15 | 10.95 | 11.36 | 11.23 | 11.71 | 11.37 | 13.01 | 13.16 | 11.95 |
| $H_2O+$ | 0.99 | 1.15 | 0.46 | 0.76 | 1.52 | 1.38 | | | | | | | | |
| $H_2O-$ | 0 | 0.19 | | | 0.24 | | | | | | | | | |
| pyrite | 1.60 | 2.16 | 2.17 | 1.95 | 1.88 | 1.83 | | | | | | | | |

| | 15 | 16 | 17 | 18 | 19 | 20 | 21 | 22 | 23 | 24 | 25 | 26 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **LOW-POTASH RHYOLITES** | | | | | | | | | | | | |
| $SiO_2$ | 70.26 | 68.36 | 70.90 | 73.60 | 72.92 | 71.38 | 70.94 | 69.72 | 70.48 | 71.16 | 63.82 | 75.22 |
| $Al_2O_3$ | 14.30 | 15.58 | 14.59 | 14.52 | 13.88 | 15.49 | | | | | | |
| $Fe_2O_3$ | 1.66 | | 0.39 | | | | | | | | | |
| FeO | | | | 0.93 | 1.59 | 0.81 | | | | | | |
| MgO | | | 0.45 | 0.90 | 1.45 | 0.72 | | | | | | |
| CaO | 0.85 | 1.74 | 1.40 | 0.92 | 2.52 | 0.84 | | | | | | |
| $Na_2O$ | 3.60 | 3.02 | 1.84 | 2.57 | 2.79 | 4.48 | 3.80 | 2.39 | 3.16 | 4.18 | 3.50 | 3.33 |
| $K_2O$ | 6.43 | 6.25 | 6.11 | 5.36 | 6.07 | 5.63 | 5.82 | 7.50 | 6.67 | 5.18 | 6.63 | 5.11 |
| $H_2O+$ | 1.29 | 2.56 | 2.23 | | | | | | | | | |
| $H_2O-$ | 0.19 | 0.60 | 0.73 | 1.89 | 1.49 | 1.59 | | | | | | |
| pyrite | 1.76 | 2.03 | 2.03 | 1.09 | 0.48 | 1.76 | | | | | | |

Analyses 1–5 and 15–17 inclusive by Sharwood, W. J. (1911, p. 734).
Analyses 6, 18, 19, 20 by Adams, E. W.
Analyses 7–14 and 21–26 inclusive by Gonyer, F. A.
* Probably about 1.00 to 1.25 high because of iron introduced by grinder.

examined under the microscope before being analyzed, to make sure that there was little or no hydrothermal alteration. No sericitized material was used.

There seems to be no relation between composition and texture of the rhyolites. About one-half of the specimens of the low-potash type are microgranular and the other half are micropoikilitic. Similarly, about one-half of the high-potash rhyolites are microgranular and one-half micropoikilitic. There is, however, some correlation between composition and physical properties of the hand specimens. When all the specimens of known high-potash composition are placed in one group and all those of known low-potash composition in another group, the high-potash specimens are

found to be darker and denser than the low-potash specimens. On the other hand, many specimens of intermediate texture and color cannot be classified with any degree of certainty.

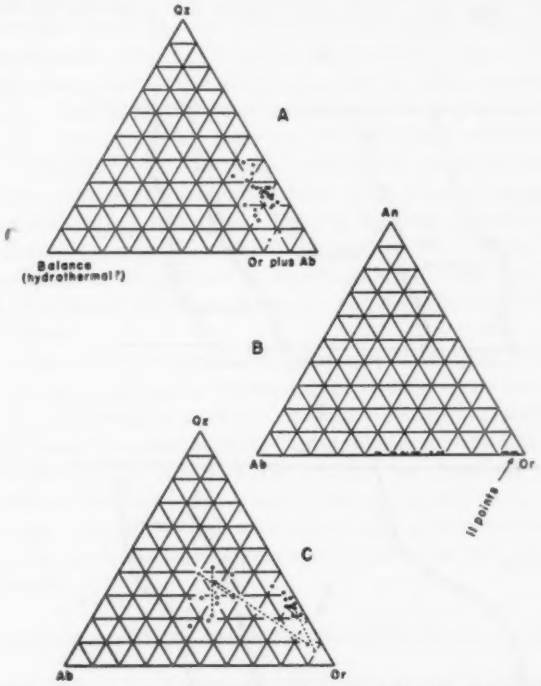



FIGURE 1.—*Ternary diagrams of normative mineral compositions of rhyolite dikes in the Homestake Mine*

A. Components quartz—(orthoclase plus albite)—balance. Microscopic study indicates that the "balance" is almost wholly hydrothermally introduced calcite and pyrite.

B. Components orthoclase-albite-anorthite. On the assumption that the "balance" of A is wholly hydrothermal, there is no anorthite. The amount of error that this assumption introduces into the diagram would be scarcely visible in the plotting.

C. Components orthoclase-albite-quartz. Crosses indicate compositions of specimens from marginal portions of dikes. The dotted lines connect these compositions with those of the central portions of the same dikes.

Not enough analysed specimens are available to make any generalizations as to age relations of the high- and low-potash types. Two specimens each from two dikes of known age relations were analysed, but all four specimens were the low-potash type. Two specimens from each of two other dikes were analysed. In this case one specimen of the earlier dike was the low-potash type, one was high-potash, and both specimens from the later dike were high-potash. These dikes were well exposed on crosscut walls, and the age relations were established beyond any reasonable doubt (Fig. 2).

The presence of high- and low-potash material in the same dike is somewhat puzzling. A possible explanation is that the high-potash material is not truly

representative of the high-potash magma but is due merely to selective precipitation of orthoclase at the dike wall because of a slight amount of assimilation of schist, orthoclase being the phase with which the magma was saturated at this stage.

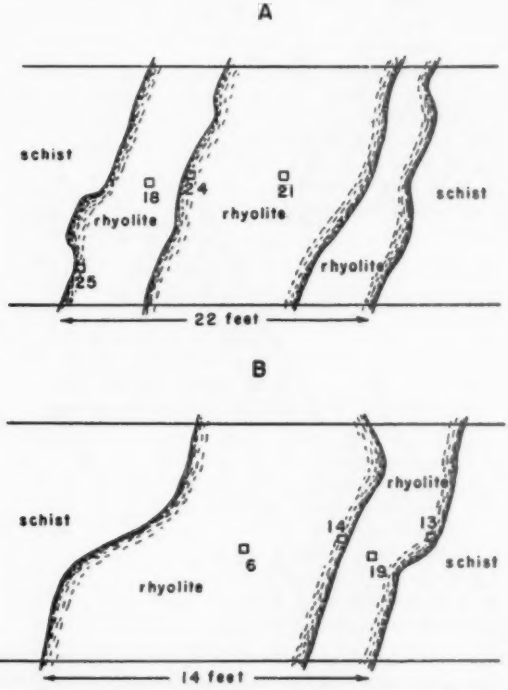



FIGURE 2.—*Intersections of rhyolite dikes in the Homestake Mine*

A. An older dike is cut through the center by a younger one. Four specimens, shown by numbers, are the low potash type.

B. A younger dike is intruded along the wall of an older one. Three specimens, Nos. 6, 14, and 13, are the high-potash type, specimen No. 19 is the low-potash type.

The numbers correspond to the numbers in Table 1.

Study of the ternary diagram of Figure 1-C shows that two of the high potash analyses from dike walls are really somewhat different from the characteristic high-potash analyses. The analysis on Figure 1-C shown by a cross near the Or corner which is connected with a low-potash dike near the center of the diagram is number 13 of Figure 2.

## OTHER OCCURRENCES OF HIGH-POTASH DIKES

In the "Lebachose" bracket of Washington's tables of chemical analyses (Washington, 1917, p. 107–108) are listed 12 analyses of rocks similar to the high-potash dikes of the Homestake Mine, but two of them are Sharwood's analyses of Homestake

dikes; hence Washington lists only 10 other known occurrences. A few more analyses have been published since the date of the tables. One analysis given by Broderick (1935, p. 536) is a duplication of material from one of the occurrences in Washington's tables. Analyses of hydrothermally altered rhyolites in the Upper Geyser Basin of Yellowstone Park, published by Fenner (1936), grade from the low-potash type to some nearly like the Homestake high-potash rhyolites. Callaghan (1937, p. 27-30) lists an analysis of a rhyolite from Lincoln County, Nevada, which has a very high potash:soda ratio and is described as being essentially a quartz-orthoclase rock. Thus there are 12 other examples of high-potash rocks of the Homestake type. Probably a few more analyses have been published that I have not seen. A study of 9 of the 12 descriptions available to me emphasizes the scarcity of this rock type. One of the analyses is from a rim on a nodule in an orbicular granite and is not a distinct rock type. Three of the descriptions specifically mention an alteration of the rock. Several of the analyses are published without any mention of the abnormal composition or any description of the rock. In all, four rhyolite flows, from Japan, Michigan, Colorado, and Nevada, are well described and are not visibly altered.

Wells (1937) lists no other examples of igneous rocks analysed in the laboratory of the U. S. Geological Survey in the years 1914 to 1936 which have potash:soda ratio equal to that of the high-potash type at Lead. The igneous rocks with nearest potash:soda ratio are two granites from Climax, Colorado with a little lower ratio than those of the Lead rhyolites. These granites are described by Butler and Vanderwilt (1933, p. 220–226) as being hydrothermally altered and containing small veins of orthoclase. Another analysis with potash:soda ratio almost as high as these granites is described as consisting of adularia and quartz, but this may be a product of ore mineralization.

## ORIGIN OF THE HIGH-POTASH RHYOLITE DIKES IN THE HOMESTAKE MINE

The low-potash dikes in the Homestake Mine are most easily explained as products of normal differentiation. They are not far from the composition of Vogt's granite eutectic (Vogt, 1931a), which has the composition: quartz, 30%; orthoclase, 28%; albite plus anorthite, 42%. Granites, porphyries, and rhyolites of this composition are so abundant that Vogt considered this to be the normal ternary eutectic of a quartz-rich magma. On the ternary diagram Or-Ab-$SiO_2$ the low-potash dikes are not far from the belt of low melting temperatures discussed by Bowen (1937), but they stand a little to one side of that belt, in the direction of higher potash content. The high-potash dikes, on the other hand, have passed far beyond this composition, to a position on the boundary line between orthoclase and leucite. Potash feldspar has taken the place of soda feldspar, without changing the quartz content. The high-potash dikes are distinguished from the granite eutectic by their high potash: soda ratio and from the feldspathoidal alkaline rocks by their high quartz content.

Bowen (1928, p. 227–233) has discussed a mechanism by which potash-rich magmas might form. The course of crystallization can be best shown on the orthoclase-albite-anorthite equilibrium diagram (Fig. 3). Any magma resulting from a normal

process of differentiation will start on the plagioclase side of the field. The course of crystallization, shown in dotted lines on the diagram, may be accompanied either by a low degree of reaction (line X-Y) or by a high degree of reaction (line X-Z). In the former instance, on reaching point Y the liquid continues in the direction of E, the

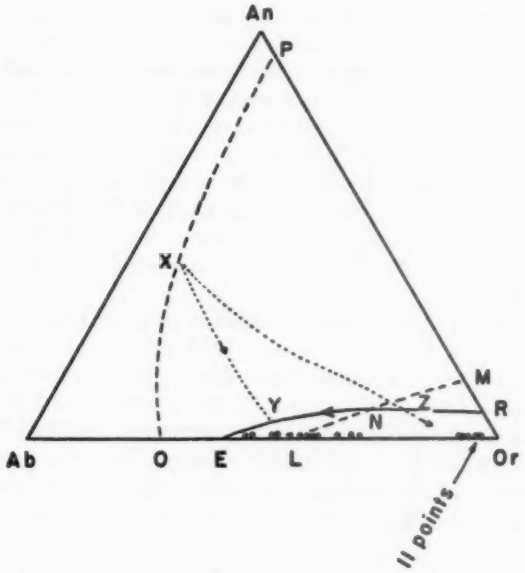



FIGURE 3.—*Ternary diagram of the system Or-An-Ab* (after Bowen, 1928)

Shows the possibility of production of potash-rich liquid of a granitic nature by fractional crystallization. Analyses of the rhyolite dikes in the Homestake Mine are plotted by dots. (*See also* Figure 1-B.)

orthoclase-albite eutectic, until it is used up. In the latter instance with high degree of reaction, however, if the course of crystallization reaches the cotectic line R-E beyond the solid solution line L-M, the orthoclase now forming has a reaction relation to the earlier-formed calcic plagioclase and to the liquid. In the normal case, reaction will take place, and the course of crystallization will again move toward E. But if for any reason, such as zoning of plagioclase or separation of liquor from crystals, reaction is prevented, a high-potash liquor will be formed. Thus an early period of strong reaction if followed by a period of hindered reaction might produce a potash-rich magma. At one stage in the process of crystallization, a potash-rich liquid may exist, which, if separated from the crystals already formed, can become a high-potash magma. This mechanism would at the same time produce rocks of high soda content, and in fact such rocks, the grorudites, do occur in the northern Black Hills.

Vogt (1905; 1931a; 1931b) has constructed a somewhat different orthoclase-albite-anorthite equilibrium, by studying the sequence of crystallization of the two kinds of feldspars, by studying what he calls the residual magma dikes with pheno-

crysts of both orthoclase and plagioclase, and by plotting analyses from all available sources, chiefly the "superior analyses" of Washington's tables. Vogt's equilibrium diagram differs from Bowen's in the orthoclase-anorthite equilibrium relations, and he makes no use of the reaction principle. But in constructing the equilibrium diagram Vogt refused to use a small number of analyses from the "superior analyses" of the tables, apparently in the belief that they were old and perhaps inaccurate. Thus the 12 analyses in the "Lebachose" bracket, including two of the Homestake dikes by Sharwood, were excluded. The analyses cannot be excluded on the basis of inaccuracy, however, because additional analyses of the Homestake rocks agree closely with the earlier ones. By excluding some analyses Vogt refused to admit the possibility that under certain unusual conditions rocks of abnormal composition can form.

Doggett (1929; Terzaghi, 1935a; 1935b) has used a statistical study of normative feldspar calculated from published analyses to support Bowen's feldspar equilibrium diagram as compared with Vogt's. She concludes, from a study of selected analyses, that most potash-rich rocks are alterations of normal magmas, and she doubts if potash-rich rocks form as products of primary crystallization. Like Vogt, she refuses to admit that unusual conditions of crystallization can produce an abnormal magma.

Another possible mode of origin of high-potash dikes is by introduction of potash in hydrothermal solutions. Vogt (1931a) explains the formation of high-potash pegmatites by this mechanism. The process of myrmekitization, whereby the potash of potash feldspars is replaced by soda and lime, forms a residual liquid rich in potash. This liquid may escape into the surrounding rocks and cause replacement of the ferromagnesian minerals by biotite and muscovite, or it may crystallize as a high-potash pegmatite. Vogt has shown that in Norway there is a close space relation between myrmekite and the high-potash pegmatites. Although these pegmatites have gone beyond the ternary granite eutectic, only about 60 per cent of the total feldspar is orthoclase, whereas in some of the high-potash dikes of the Homestake Mine about 95 per cent of the total feldspar is orthoclase. The high-potash pegmatites may be an intermediate step between the granite eutectic and the high-potash dikes, but they have gone only a short part of the way. Moreover, myrmekite has not been described in the Tertiary igneous rocks of the northern Black Hills.

If the process of myrmekitization can convert potash feldspars to soda-lime feldspars, the reverse process may convert normal granites into high-potash rocks. Fenner (1936) found that hydrothermal solutions had substituted orthoclase for plagioclase feldspars in the rhyolites of the Upper Geyser Basin in Yellowstone Park. His conclusions as to the origin of the high-potash rhyolites of Yellowstone seem to be supported by the evidence in that occurrence. The only support I can find for a hydrothermal origin of the high-potash dikes in the Homestake Mine is the presence of both high-potash and low-potash material in the same dike (Fig. 2), and this can have a different explanation. On the other hand, a hydrothermal origin is strongly opposed by two relationships: the complete lack of gradational types; and the relatively slight alteration in the schist wall rocks. Sericite, not orthoclase, would be expected to result from the hydrothermal replacement by potash-rich solutions.

Chapman (1939, p. 167–170) believes that rocks in New Hampshire of composition equivalent to quartz monzonite and granite have been converted from quartz diorite and granodiorite by potash-rich solutions rising from the magma reservoir below through already crystallized material. The evidence for hydrothermal replacement in these New Hampshire occurrences is convincing, but similar evidence is wholly lacking for the high-potash dikes of the Lead area.

Assimilation of wall rocks has been cited as the origin of some of the low-silica alkaline rocks, and there is evidence that the Homestake dikes have assimilated or otherwise removed at least a small part of the wall rocks. But there seems to be no possibility that assimilation is the origin of the high potash content. The field relations and mineral and chemical composition of the dike edges are proof of some assimilation. Some highly irregular contacts between schist and dike can only be explained as a removal of the schist by the dike magma, and a concentration of orthoclase at some dike edges is evidence in support of reactive assimilation of schist. Most of the schists have lower silica content than do the rhyolites, but quartzites have higher silica content. Of the six pairs of analyses of specimens from dike centers and dike edges (Fig. 1), five were adjacent to schist and one was adjacent to quartzite. The rhyolite lost silica in four of the five pairs of analyses adjacent to schist, and in one pair there was no change, but the rhyolite gained silica in the pair of analyses adjacent to quartzite. On the other hand, the field relations will not permit a great amount of assimilation of wall rock. Moreover, although some potash can be gained by selective assimilation of the biotite of the schists, there is no means of reducing soda almost to zero while silica remains about constant. Furthermore, variation diagrams show that no reasonable amount of any natural substance can be added to the low-potash dikes to produce the high-potash dikes.

Three possible modes of origin of the high-potash rhyolite dikes in the Homestake mine have been suggested: (1) normal differentiation with strong reaction, followed by hindered reaction, probably by a separation of liquid from crystals; (2) hydrothermal introduction of potash, either as a residual liquid akin to a pegmatite solution or as a potash-rich solution capable of converting soda-lime feldspars to potash feldspars; and (3) selective assimilation of potash-rich wall rocks. The first mode of origin seems to be theoretically sound and is not opposed by the observed facts. It has the merit of operating only accidentally and occasionally. The two hydrothermal theories can be supported, but lack of gradational types and lack of schist alteration are objections to it. The theory of assimilation probably cannot be supported as a cause of the high-potash content of the Homestake dikes.

## REFERENCES CITED

Bowen, N. L. (1928) *The evolution of the igneous rocks*, Princeton Univ. Press, 332 p.

——— (1937) *Recent high-temperature research on silicates and its significance in igneous geology*, Am. Jour. Sci., 5th ser., vol. 33, p. 1–21.

Broderick, T. M. (1935) *Differentiation in lavas of the Michigan Keweenawan*, Geol. Soc. Am., Bull., vol. 46, p. 503–558.

Butler, B. S., and Vanderwilt, J. W. (1933) *The Climax molybdenum deposit, Colorado*, U. S. Geol. Survey, Bull. 846, p. 195–237.

Callaghan, Eugene (1937) *Geology of the Delamar District, Lincoln County, Nevada*, Univ. Nevada, Bull., vol. 31, 72 p.

Chapman, C. A. (1939) *Geology of the Mascoma Quadrangle, New Hampshire*, Geol. Soc. Am., Bull., vol. 50, p. 127–180.

Darton, N. H. (1912) *Volcanic action in the Black Hills of South Dakota*, Science, new ser., vol. 36, p. 602–603.

Darton, N. H., and Paige, Sidney (1925) *Description of the Central Black Hills*, U. S. Geol. Survey, Geol. Atlas, Folio no. 219.

Doggett, R. A. (1929) *The orthoclase-plagioclase equilibrium diagram*, Jour. Geol., vol. 37, p. 712–716.

Fenner, C. N. (1936) *Bore-hole investigation in Yellowstone*, Jour. Geol., vol. 44, p. 225–315.

Irving, J. D. (1899) *A contribution to the geology of the northern Black Hills*, N. Y. Acad. Sci., Ann., vol. 12.

Jaggar, T. A. (1901) *The laccoliths of the Black Hills*, U. S. Geol. Survey, 21st Ann. Rept., pt. 3, p. 163–303.

Ludlow, William (1875) *Report of a reconnaissance of the Black Hills of Dakota made in the summer of 1874*, U. S. Army, Eng. Dept., 121 p.

Newton, Henry, and Jenney, W. P. (1880) *Report on the geology and resources of the Black Hills region, Dakota*, U. S. Geog. Geol. Survey Rocky Mtn. Region, 566 p.

Sharwood, W. J. (1911) *Analyses of some rocks and minerals from the Homestake Mine, Lead, South Dakota*, Econ. Geol., vol. 6, p. 729–789.

Terzaghi, R. D. (1935a) *The origin of the potash-rich rocks*, Am. Jour. Sci., 5th ser., vol. 29, p. 369–380.

——— (1935b) *The origin of the potash-rich rocks*, Am. Jour. Sci., 5th ser., vol. 30, p. 141–142.

Vogt, J. H. L. (1905) *Phys.-Chem. Getsetze*, Min. pet. Mitt., vol. 24.

——— (1931a) *The physical chemistry of the igneous rocks*, Norske Vidensk.-akad. Oslo, 1930.

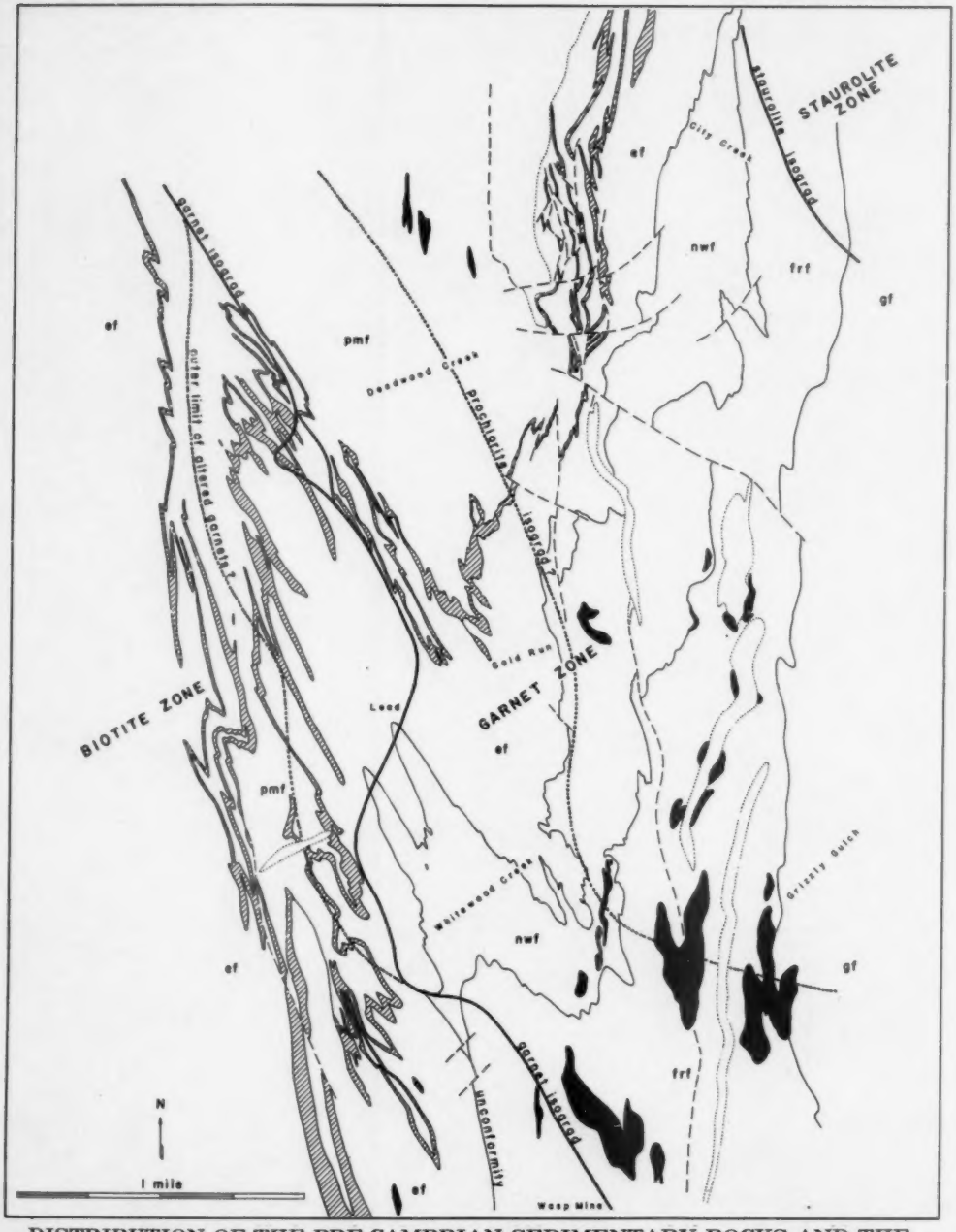
——— (1931b) *On the terms eutectic, cotectic, peritectic, anchi-eutectic, anchi-cotectic, etc., and their importance in petrogenesis*, Jour. Geol., vol. 39, p. 401–431.

Washington, H. S. (1917) *Chemical analyses of igneous rocks*, U. S. Geol. Survey, Prof. Paper 99, 1201 p.

Wells, R. C. (1937) *Analyses of rocks and minerals from the laboratory of the United States Geological Survey, 1914–36*, U. S. Geol. Survey, Bull. 878, 134 p.

California Institute of Technology, Pasadena 4, California

Manuscript Received by the Secretary of the Society, August 19, 1947




## DISTRIBUTION OF THE PRE-CAMBRIAN SEDIMENTARY ROCKS AND THE METAMORPHIC ZONES IN THOSE ROCKS IN THE LEAD DISTRICT

From mapping on scale of 1 inch to 200 feet, except a small part on scale of 1 inch to 1000 feet. Large amphibolites are in solid black. Large Tertiary intrusive bodies are dotted. Omitted Cambrian strata, small amphibolites, small Tertiary intrusive bodies (including sills in the Cambrian), and Tertiary gravel deposits. Faults are shown by dashed lines. Poorman formation, pmf; Ellison formation, ef; Northwestern formation, nwf; Flag Rock formation, frf; Grizzly formation, gf; Homestake formation, cross hatched.

BOARD

S

Abstr
Introd
Strati
G
S

I
C
Metam
G
H
G
S
P
P
P
P
M
C
P
C
R
Refere

Figure
1. S

Plate
1. D

2. P
3. P

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 59, PP. 941-976, 1 FIG., 3 PLS. SEPTEMBER 1948

# STRATIGRAPHY AND METAMORPHISM IN A PART OF THE NORTHERN BLACK HILLS AND THE HOMESTAKE MINE, LEAD, SOUTH DAKOTA

BY JAMES A. NOBLE AND JAMES O. HARDER

## CONTENTS

| | Page |
|---|---|
| Abstract | 942 |
| Introduction | 942 |
| Stratigraphy | 943 |
| General | 943 |
| Sedimentary rocks | 944 |
| Metamorphic zoning | 944 |
| Poorman formation | 945 |
| Homestake formation | 946 |
| Ellison formation | 947 |
| Northwestern formation | 948 |
| Flag Rock formation | 949 |
| Grizzly formation | 952 |
| Cambrian Deadwood formation | 952 |
| Gravel deposits | 953 |
| Igneous rocks | 954 |
| Correlations | 954 |
| Metamorphism | 955 |
| General | 955 |
| Biotite zone in the argillaceous rocks | 956 |
| Garnet zone in the argillaceous rocks | 956 |
| Staurolite zone in the argillaceous rocks | 969 |
| Progressive metamorphism in quartzites | 960 |
| Progressive metamorphism of lime-bearing iron carbonate rocks | 962 |
| Progressive metamorphism of lime-free iron carbonate rocks | 963 |
| Progressive metamorphism of amphibolites | 965 |
| Mineral facies | 966 |
| Changes of composition during metamorphism | 968 |
| Probable shapes of the isothermal surfaces | 969 |
| Cause of the progressive metamorphism | 970 |
| Retrogressive metamorphism | 971 |
| References cited | 971 |

## ILLUSTRATIONS

| Figure | Page |
|---|---|
| 1. Stratigraphic column of the pre-Cambrian rocks | 944 |

| Plate | Facing page |
|---|---|
| 1. Distribution of the pre-Cambrian sedimentary rocks and the metamorphic zones in those rocks in the Lead district | 941 |
| 2. Photomicrographs of typical sections of metamorphic rocks from the Lead district | 974 |
| 3. Photomicrographs of special features in the metamorphic rocks of the Lead district | 975 |

## ABSTRACT

The pre-Cambrian rocks of the northern Black Hills which have been mapped and studied by the staff of the Homestake Mining Company comprise sedimentary formations with a total stratigraphic thickness of about 20,000 feet. The oldest formations, the Poorman and Homestake, are dominantly ankeritic carbonate and iron-magnesium carbonate respectively. The younger formations, the Ellison, Northwestern, Grizzly, and Flag Rock, are dominantly argillaceous rocks with some quartzites and other rock types. Pre-Cambrian intrusive rocks, originally gabbroic, cut the pre-Cambrian sedimentary rocks; and the whole succession is overlain unconformably by the Cambrian Deadwood formation, and everything is cut by Tertiary intrusive rocks.

Regional metamorphism of pre-Cambrian age, presumably related to an undisclosed granitic intrusion somewhere northeast of the district, has developed metamorphic zones of the Scottish Highlands types. Three metamorphic zones are developed in the argillaceous rocks: a biotite zone; a garnet zone; and a staurolite zone. The iron-magnesium carbonate, sideroplesite, the essential mineral of the Homestake formation, is converted to the iron-magnesium silicate, cummingtonite, during this metamorphism. Some of the mineralogic changes that took place during this regional metamorphism suggest that there were changes in the bulk composition of the rocks whereby potash migrated outward from zones of higher metamorphic grade ahead of the advancing isothermal surfaces.


## INTRODUCTION

This paper gives the results of studies of the stratigraphy and metamorphism of the pre-Cambrian rocks in the vicinity of the Homestake Mine, in the northern part of the Black Hills of South Dakota. The field and laboratory work on which these studies are based was carried out at intervals over about 12 years, with the assistance at one time or another from the following present or former colleagues of the authors: C. N. Kravig; T. A. Dodge; R. G. Wayland; A. L. Slaughter; G. C. Mathisrud; T. M. Rizzi; and D. C. Lavier. The manuscript has been read critically by Professors E. S. Larsen and M. P. Billings of Harvard University, G. M. Schwartz and J. W. Gruner of the University of Minnesota, and Dr. T. A. Dodge of Tucson, Arizona, to all of whom we are greatly indebted for much constructive criticism. For opportunity to carry on this scientific work and to publish the results we are indebted to officials of the Homestake Mining Company, particularly to Dr. D. H. McLaughlin, president of that company. Last, but not least, sincere thanks are owed to the many visitors who have spent time with us, have seen our data in the field, and have agreed or disagreed with our conclusions.

Although this is the first statement of a zonal pattern of metamorphism in the northern Black Hills, it is by no means the first paper on the stratigraphy and metamorphism of the pre-Cambrian rocks in that area. The most important earlier papers, listed in the order of their publication, are those by Hosted and Wright (1923), Paige (1924), Darton and Paige (1925), McLaughlin (1931), and Gustafson (1933).

The broad features of the geology of the Black Hills have been discussed adequately by Darton and Paige (1925). A large oval area of pre-Cambrian rocks is flanked on the northwest side by two smaller areas of pre-Cambrian rocks, and all three are surrounded by outward-dipping Paleozoic, Mesozoic, and Tertiary sedimentary rocks. The pre-Cambrian sedimentary rocks are almost everywhere highly folded and converted into phyllites or schists. Large and small masses of amphibolite represent

medium-basic igneous rocks intruded probably near the end of the period of folding. In the southern part of the large area, the pre-Cambrian rocks are intruded and deformed by several masses of coarse granite and by many pegmatite dikes and pipes. In addition, there are small masses of gneissic granite in the northeast part of the main schist area on Little Elk Creek, some pegmatite pipes and dikes in one of the smaller schist areas at Tinton, and a few small pegmatite dikes in a patch of schist adjacent to the second schist area, on Whitewood Peak, 5 miles northeast of Lead. All these granites and pegmatites show close similarity in the heavy mineral suite and probably are co-magmatic (Marsden, 1933). Tertiary granitic, monzonitic, and syenitic porphyries cut and deform the schists and Paleozoic and Mesozoic rocks in the northern Hills in the extreme northern part of the main schist area, in each of the smaller schist areas, and to considerable distances in the Paleozoic and Mesozoic sedimentary rocks to the northeast, north, and northwest of those areas. No Tertiary intrusive rocks crop out in the southern Hills.

The geologic history of the Lead district can be briefly summarized as follows: In pre-Cambrian time, a thick series of mudstones, with some other types of rocks, were formed and later highly folded and transformed by progressive regional metamorphism. Igneous rocks, intruded probably near the end of the period of folding, were also metamorphosed to amphibolites. Extreme compression, resulting in nearly plastic flow of the rocks, produced tenuous, closed anticlines and synclines; but subsequently new forces, or changing orientations of the original structures, caused refolding and deformation of these isoclinal folds. Most of the metamorphic changes described in this paper took place in pre-Cambrian time during or following the first period of folding. The age of the second period of folding has not been determined, and it may be much later. After profound erosion, to a stage of peneplanation, sediments of Paleozoic and Mesozoic age accumulated. In early Tertiary time, igneous rocks in great abundance and considerable variety were injected into the schists and the later sedimentary rocks, their intrusion being coincident with a period of orogeny which lifted the area to practically its present form. Some of the smaller intrusions, by forcible injection, have deformed the shist wall-rocks for distances of a few feet, and some of the larger intrusions have had even greater deformational effects. Closely related to the Tertiary igneous activity was a gold mineralization accompanied by pyrite, calcite, sericite, and quartz, with insignificant amounts of silver, lead, zinc, and tungsten. The ores of several small districts surrounding the Lead district were formed in this period. Whether the much more important deposits of the Homestake Mine were formed in this period or in pre-Cambrian time is still debated. This question, as well as many others related to Tertiary intrusion, structure, and ore mineralization, will be discussed in separate papers.

## STRATIGRAPHY

### GENERAL

The column of stratigraphy and igneous intrusion established by the study of the rocks of the Lead district is as follows:

Tertiary, Quarterary, and Recent gravel deposits (at least 90 feet; probably much more)
—— unconformity ——
Tertiary intrusive rocks

Cambrian Deadwood formation (300 to 500 feet)
------- unconformity -------
Pre-Cambrian intrusive rocks
Pre-Cambrian sedimentary rocks
Grizzly formation (possibly 3000 feet or more)
Flag Rock formation (possibly 5000 feet)
------- unconformity -------
Northwestern formation (possibly 4000 feet)
Ellison formation (3000 to 5000 feet)
Homestake formation (200 to 300 feet)
Poorman formation (possibly 2000 feet or more)

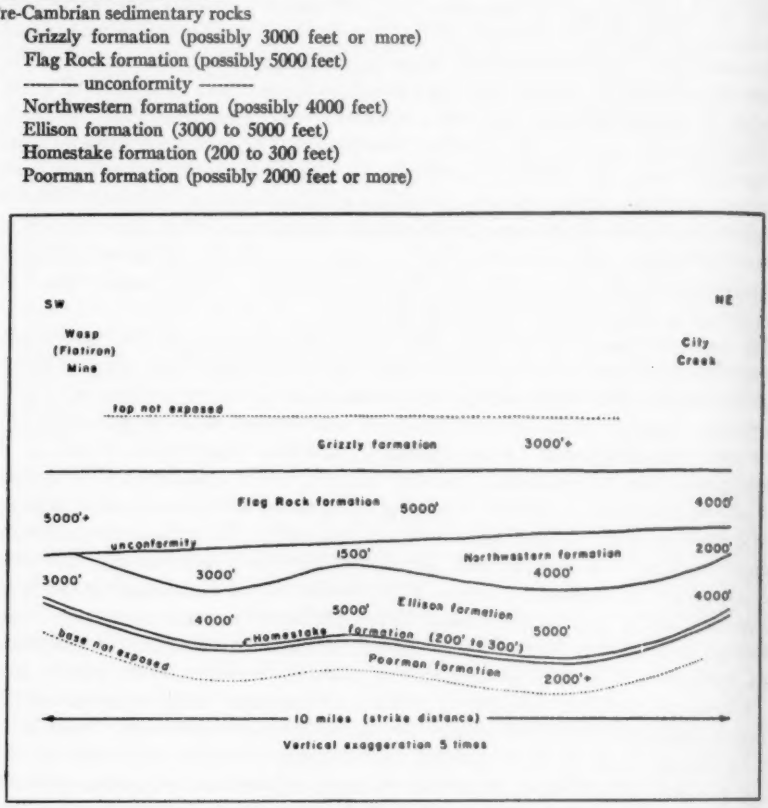



FIGURE 1.—*Stratigraphic column of the pre-Cambrian rocks in the Lead district*

Figure 1 shows the pre-Cambrian sedimentary rocks. The thicknesses of these pre-Cambrian rocks are estimated on the assumption that the present shapes of the folds in the rocks are a result of shear folding rather than flexure folding. The thicknesses are for the rocks in the present state of compaction, not for unmetamorphosed rocks.

## SEDIMENTARY ROCKS

*Metamorphic zoning.*—Some mention must be made of the metamorphic zoning, to be discussed more fully in the second section. The pre-Cambrian sedimentary rocks of the Lead district show three well-defined zones of metamorphism, with increase of grade to the northeast, the isograd or isothermal lines running in a north-northwest direction (Pl. 1). In this section the characteristics of the rocks in the outermost,

lowest grade, biotite zone will be described; and in a succeeding section the progressive metamorphism through all the metamorphic zones.

*Poorman formation.*—The characteristic material of the Poorman formation is gray phyllite. The most common color is dark gray, in a few places becoming very dark or even sooty black; but lighter shades of gray are not uncommon. Green and greenish gray, brown, reddish gray, and red occur in places. A very small portion of the formation has grain fine enough and schistosity good enough to be described as slate, and there is also a very small portion somewhat coarser than average and properly described as schist; but for most of the formation the term phyllite is most appropriate. Much of the material is finely laminated, and on some outcrops the lamination is emphasized by differential solution on a small scale. This type of weathering, together with the redeposition of networks of tiny calcite veinlets, points to a content of carbonate in the Poorman but gives no hint of the predominance of carbonate shown by some of the material under the microscope. The formation almost everywhere shows enough bedding to disclose structure; although massive, structureless outcrops occur, they are very rare. Laminae range from about 1 mm. to about 5 mm. thick. In addition to this lamination there is an erratic distribution of banding, which, within the mine area, is most common in the uppermost 100 feet of the formation. In the mine mapping, we sometimes have treated this banded portion as a separate formation, under the name of De Smet formation, but throughout the district this banded portion has no constant position; it may be lacking or may occur at one or at several different positions. The De Smet has therefore been dropped as a separate formation. The banding is a type of small-scale differentiation or segregation strictly parallel to original bedding, individual bands ranging from about 3 mm. to about 2 cm. thick. It is partly produced by differing amounts of graphite and partly by differences in color due to segregation of metamorphic minerals. In many places pods or bands of recrystallized quartz occur, usually dull gray, probably originally pods and bands of chert. Cubes of pyrite are not uncommon in the dark-gray or black phyllites. A few narrow lenticular beds of brown sideroplesite schist are about 200 feet from the top of the formation in some areas. These are not persistent, however, and have little value as horizon markers.

Under the microscope, the dark, laminated or massive phyllites of the Poorman formation typically show only quartz, muscovite of sericitic habit, graphite, oxides of iron, and usually ankerite. Phlogopite has been noted in a few sections but is not common. Usually quartz is the most abundant constituent, muscovite next, and graphite next; ankerite may be absent or may be dominant. The muscovite has good dimensional orientation parallel to bedding, and quartz and ankerite have some dimensional orientation; but with increase in amount of carbonate, the dimensional orientation of all minerals becomes poor. Neither biotite nor chlorite is found in this type of material from the biotite zone; but all the banded phyllite or schist (as distinct from the finely laminated or massive phyllite) has biotite instead of muscovite. In this material, quartz, biotite, graphite, tourmaline, and iron oxides are the only constituents. The most abundant constituent in most sections is biotite. Chlorite is not found except near amphibolite masses or quartz veins. The narrow, nonpersistent beds of sideroplesite schist contain quartz, sericite, and sideroplesite.

The primary carbonate of the Poorman formation has hitherto been called dolomite (Paige, 1924, p. 15–18; Darton and Paige, 1925, p. 3; Gustafson, 1930, p. 18, 1933, p. 129; McLaughlin, 1931, p. 325), but strictly speaking it is ankerite. Several measurements of $\omega$ give 1.697, and a few go as high as 1.710; none was recorded lower than 1.695. Chemical analyses of specimens of the first type show the carbonate to contain approximately 15 per cent $CaFeC_2O_6$, 85 per cent $CaMgC_2O_6$, which is probably too far from pure $CaMgC_2O_6$ to be called dolomite. The second type of carbonate, with $\omega = 1.710$, by analysis has approximately 45 per cent $CaFeC_2O_6$; this obviously cannot be called dolomite.

The ankerite of the Poorman formation is probably sedimentary. The mode of origin of a sediment of this type is not known.

The Poorman formation in large outcrops is not easily confused with other formations, but small outcrops may be confused with phyllites of the Ellison, Northwestern, or Flag Rock formations. Ankerite, though not universal, is fairly common and fairly distinctive; but a little ankerite occurs in some of the Ellison phyllite. There seem to be no beds of unusual character sufficiently persistent for horizon markers in structural studies.

The Poorman formation is estimated to be at least 2000 feet thick, but the bottom of the formation has not been reached by the mapping. This estimate may, however, be high due to undetected duplication by folding.

*Homestake formation.*—The Homestake formation, in the outer metamorphic zone, is sideroplesite-quartz schist (sideroplesite being the iron-magnesium carbonate $FeCO_3$,$MgCO_3$). Most outcrops are red, reddish brown, or brown, due to the iron content; but greenish brown and green, due to the presence of chlorite, are not uncommon. Gray is rare on outcrops, but some mine exposures are gray or dark gray. Pods and seams of light-gray to white recrystallized quartz, probably originally chert, are characteristic; in fact, are rarely lacking. In only a few places are the sideroplesite grains coarse enough to be readily visible to the eye; most grains are well under 1 millimeter, except in zones of hydrothermal activity. Banding due to different amounts of graphite occurs in a few places; some small lenses of laminated gray schist have been grouped with the Homestake formation on the surface maps, but these may be infolded pieces of other formations. Small veins of white quartz occur abundantly in the recrystallized quartz bands and in sideroplesite schist.

The Homestake formation contains sideroplesite, quartz, biotite, chlorite, and usually a little graphite. Dimensional orientation of mineral grains is good where mica is abundant but becomes poor as the carbonate content increases. There is usually some coarse biotite in a random or perhaps criss-cross orientation. Chlorite plates of the same size as those of biotite and in the same mode of occurrence are probably unit pseudomorphs of daphnite after the biotite and probably are hydrothermal. In other sections biotite may have formed at the expense of primary chlorite, but the relationship is not certain. Thin sections of the small pods and bands of recrystallized quartz show a mosaic of polygonal quartz grains, with small amounts of sideroplesite. In one section there is a little biotite, but the absence of micas or other aluminous minerals in the others supports the belief that these pods and bands were originally chert, not accumulations of clastic quartz grains.

Sideroplesite schist, with abundant pods of recrystallized quartz, characterizes the Homestake formation and outside that formation occurs only in one or two narrow beds in the Poorman formation, one bed in the Ellison formation, and two or three narrow beds in the Flag Rock formation. Heavily ironstained outcrops suggest the possibility of occurrence of sideroplesite schist, but of course there are other sources of iron. Where there is enough chlorite to give a greenish color, the formation is not readily distinguished from some phases of the Poorman or Ellison formations. The pods and bands of recrystallized quartz also occur in other formations.

The Homestake formation was probably 200 to 300 feet thick originally, but it has suffered great distortion during folding, so that now it is absent in places and in other places is several hundred feet thick.

The mode of origin of such a sediment as the Homestake formation is a problem that has been discussed in other regions, notably in the Lake Superior region (Van Hise and Leith, 1911; Gruner, 1922; Leith, Lund and Leith, 1935; Pettijohn, 1943; Bruce, 1945). The small amount of additional data from the Lead district probably can add very little to the solution of this problem. The uniform character of the Homestake formation in several miles of outcrop and many scores of thin sections opposes any theory of replacement of lime by iron; the sideroplesite must be explained as a sediment. Whatever the conditions leading to the accumulation of nearly pure sideroplesite (iron-magnesium carbonate) and nearly pure chert, those conditions were repeated several times, because material identical with that of the Homestake formation occurs in narrow beds and lenses in the Poorman formation, in the Ellison formation, and in the Flag Rock formation. Apparently the sideoplesite-chert combination could form almost as readily as could limestone or dolomite in more recent times. The hypothesis of a derivation of iron from igneous sources receives no help from the Black Hills area; no igneous activity is known here until much later. An interesting feature is the stratigraphic sequence, which in the Lead district is ankerite phyllite, then iron formation (sideroplesite), then quartzite, then slates and phyllites; and this sequence is nearly duplicated at Sunrise, Wyoming, where dolomite is followed by iron formation, and that by phyllites. The normal sequence in the Lake Superior region is somewhat different, however; conglomerate is followed by quartzite, that is followed by iron formation, and that in turn is followed by slates (Van Hise and Leith, 1911 p. 501).

*Ellison formation.*—The Ellison formation consists of phyllites and schists, with a considerable amount of dark quartzites; but in part of the district the middle portion of the formation has no quartzites, and the formation can be subdivided into three members: a lower member of quartzites and phyllites, about 1200 feet thick; a middle member of phyllites, about 1500 feet thick; and an upper member of quartzites and phyllites, between 1200 and 2000 feet thick. The quartzites of the two members are not sufficiently distinct to permit identification in isolated outcrops; they can be distinguished only by position. The top of the uppermost quartzite bed is taken to be the top of the formation. Since the quartzite beds are not known to be persistent, this may mean that the top of the formation as defined is not a constant horizon, but actually the variation appears to be small.

The phyllites of the Ellison formation are light gray, gray, and dark gray; in places

green gray, brown, or reddish gray; and in a very few places almost black. Iron stains are rare; but one narrow lens of sideroplesite schist occurs in the upper quartzite member. The material in a few places is best described as slate, in a few others as schist. Lamination and banding are not as common as in the Poorman formation, but both graphitic banding and segregation banding occur in a few places, always parallel to bedding. Lenses and bands of recrystallized quartz, probably originally chert, occur in a few places. A sandy texture is not uncommon in the phyllites, and some sandy phyllites are essentially micaceous quartzites. The massive quartzite beds range from about 6 inches to about 50 feet thick, but most are 2 to 10 feet thick. The lateral persistence of single beds probably is measured in hundreds of feet. Most of the quartzites are dark gray, nearly black, but some are light gray, and a few are reddish gray. Copper-red stains occur on some outcrops. Most quartzites are dense and massive, but some are micaceous and somewhat sandy. Individual quartz grains are distinguishable by eye in some outcrops; in these, well-rounded quartz grains up to 2 millimeters in diameter can be seen. There is no semblance of bedding, cross bedding, or other primary structural features in these quartzites.

Under the microscope, the Ellison phyllites consist of quartz and muscovite, in some places also biotite, in other places a little ankerite, and in still other places a few grains of tourmaline and sphene. Chlorite is not present except as a hydrothermal product. The mineral grains show very good dimensional orientation parallel to the bedding. The quartzites have, in addition to quartz, a small amount of graphite, iron oxides, sericitic muscovite, and in places a little carbonate (ankerite?). The quartz shows a mortar texture of large grains set in a groundmass of small ones; but the original clastic grains were still larger, and the outlines (most of them very well rounded) are visible in most specimens. The single bed of sideroplesite schist contains sideroplesite, and a little biotite, quartz, and chlorite (probably replacing biotite); thus it is not distinguishable microscopically from similar rocks in the Homestake formation.

The Ellison formation throughout most of the district is probably about 4000 or 5000 feet thick, allowing for duplication by folding; but in the southern part it is probably a little thinner, about 3000 feet.

*Northwestern formation.*—The Northwestern formation is almost absent from the outermost metamorphic zone, the biotite zone, as it is cut out by the unconformity at the base of the Flag Rock formation; hence its characteristics are taken from the outer part of the garnet zone, where small garnets occur in only a few outcrops. The formation consists wholly of phyllite and schist and a little slate. Most of the formation shows no lamination or bedding and is structureless; but some is laminated, and a small portion shows prominent, widely spaced black graphitic banding. The only marker suitable for disclosing structure is one bed of sandy, yellow sericitic schist, almost a micaceous sandstone; but this extends for only a few hundred feet near the top and is cut off at each end by the unconformity at the base of the Flag Rock formation.

Microscopic study shows the phyllites of the Northwestern formation to consist of quartz, sericitic muscovite, usually biotite, and usually a little tourmaline and sphene. Dimensional orientation of mineral grains is very good. Under the microscope the bed of yellow schist shows no difference from the rest of the formation.

The Northwestern formation is about 4000 feet thick in the center of the district, but the absence of good bedding makes it difficult to estimate the amount of repetition by folding. To the north, the formation narrows to about 2000 feet; to the south it disappears entirely. The original top of the formation is not exposed, and the variations in thickness are caused by erosion before deposition of the overlying Flag Rock formation.

*Flag Rock formation.*—The Flag Rock formation is difficult to describe because of its heterogeneous character and also because outcrops are not abundant in the outermost metamorphic zone. The description includes some of the formation in the outer part of the next metamorphic zone, the biotite zone. The most abundant material in the Flag Rock formation is light-gray sericitic phyllite or schist. Other colors are greenish gray and green; but reds and browns are almost entirely absent. Some of this phyllite has fine laminations, and some shows banding due to segregation of minerals and to concentration of graphite. The Flag Rock formation also has a considerable amount of soft sooty-black schist or phyllite; in this, pyrite is abundant and so widely distributed (even in the Rochford district) as to suggest a syngenetic origin for the iron and sulphur. A similar mode of origin of pyrite in graphitic rocks of the Archean of the Canadian Shield has been suggested by Pettijohn (1943). These pyritic black schists are stained yellow and white on weathered surfaces. Another conspicuous type of material in the Flag Rock formation is quartzite, which in this formation has everywhere a streaked appearance due to uneven distribution of graphite, strung out in bands and shreds through the otherwise white or light-gray rock. These quartzites are readily distinguished from the evenly dark quartzites of the Ellison formation. They range from about 2 feet to about 20 feet thick. They are not persistent on the strike and are confined to no particular horizons, but they probably do not occur in the uppermost 2000 feet of the formation. Reasons are given below for believing that these quartzites probably were originally thick beds of chert. Two or three narrow nonpersistent beds of brown sideroplesite schist (converted to cummingtonite in the garnet zone) occur in the lower half of the formation; these have pods of recrystallized quartz like that of the Homestake formation, and in fact they are not distinguishable from the Homestake formation except by thickness (they are only 5 or 10 feet thick) and by position. A few beds of yellow or nearly white sandy schist or micaceous sandstone occur near the base of the formation. The beds of streaked quartzite, sideroplesite schist, and yellow sandy schist serve as horizon markers over distances of a few hundred feet but no farther.

A special feature of the Flag Rock formation, known as the Iron Dike, forms conspicuous outcrops on hilltops to the east and northeast of Lead. This consists of dense, siliceous, jaspery material, for the most part without visible bedding but in places with faint bedding. Colors range from yellow through brown and reddish brown to black, and there is also some spongy black material, full of irregular cavities. The Iron Dike is clearly an alteration product of parts of the Flag Rock formation, and silicification is the chief process in the alteration. No great amount of iron is needed to explain the colors. A series of specimens covering a transition from black graphitic and pyritic schist to dense black and brown Iron Dike material shows a progressive removal of iron and an introduction of fine-grained quartz. In the area

near the Yates shaft, the outcrops of Iron Dike show some obscure bedding and some incompletely altered remnants of streaked graphitic quartzites. The horizon of these outcrops at a distance of only 1000 feet along the strike is mostly graphitic schists with some narrow beds of quartzite and cummingtonite schist. Other outcrops of the Iron Dike northeast of Flag Rocks are identical with those just described but are not at the same horizon; one is almost at the base of the formation, another is at 600 to 900 feet above the base, and another is near the top. These occurrences, only on hilltops, never in valleys, indicate that the Iron Dike is a product of weathering during some earlier erosion period. There are two possible erosion periods: one, a mature stage of erosion the surface of which still occupies much of the higher ground between the more recent youthful valleys; and the other, the erosion period preceding the deposition of the Upper Cambrian Deadwood formation. In the Lead district, the mature erosion surface and the base of the Deadwood formation so nearly coincide that the weathering effects cannot be distingushed. In the Rochford district, however, the mature erosion period cut well below the Deadwood formation over large areas; and in that district the Iron Dike is found only near the base of the Deadwood. Moreover, in a few places in the Lead district there are pebbles in the basal conglomerate of the Deadwood formation that closely resemble jasper from the Iron Dike. Probably, therefore, the Iron Dike is a product of weathering of different parts of the Flag Rock formation during the erosion period preceding deposition of the Upper Cambrian Deadwood formation.

Under the microscope, the light-gray phyllite and schist of the Flag Rock formation have quartz, muscovite, graphite, iron oxides, and a little tourmaline. The soft sooty black schist for the most part has only quartz and graphite, in places also muscovite; and specimens from below the zone of oxidation have pyrite cubes, which are represented by small holes in the oxidized specimens. The streaked graphitic quartzites consist mainly of quartz; in some, there is a little graphite, muscovite, iron oxides, and tourmaline, but in others there is nothing but quartz and graphite. Moreover, there is complete absence of outlines of original clastic grains, whereas most specimens of Ellison quartzites in all but the highest metamorphic zones show the outlines of clastic grains, marked by borders of graphite and iron oxide. The absence of argillaceous material in some specimens, together with the absence of outlines of clastic grains in all specimens, points to chert, rather than sands and silts, as the primary substance of these quartzites. The sideroplesite schist consists of quartz, sideroplesite, biotite, chlorite, and graphite. The yellow sandy schist contains only quartz, a little muscovite, and a little limonite. In all the material, dimensional orientation of mineral grains is good where micas are fairly abundant but is poor where quartz or sideroplesite make up most of the section.

The Flag Rock formation is probably about 5000 feet thick, but there may be some duplication by folding.

The unconformable relation between the Flag Rock formation and the underlying Northwestern formation is not readily seen. Although an approximate thickness of 4000 feet of Northwestern formation is entirely cut out in a strike distance of about 12,000 feet, there is no point at which discordance of strikes can be detected. In other regions it has been found that in highly metamorphosed rocks discordance of

bedding has been obliterated by the effects of metamorphism, but here a better explanation may be that the discordance is uniform and too small to observe. This unconformity extends from the north to the south edges of the map (Pl. 1). The Northwestern formation is possibly 4000 feet thick east of Lead, not far from the center of the map; it is only about 2000 feet thick on the north edge of the map; and it disappears completely on the south edge. Reconnaissance mapping has also shown that to the northwest from the map area the Northwestern formation differs in thickness from place to place and is sometimes lacking. Moreover, in the Rochford district, 20 miles south of Lead, the Northwestern formation is missing, and the Flag Rock formation cuts across the Ellison formation and in part of the district rests on the Homestake formation.

The sole evidence for this unconformity in the Lead district is the thinning and disappearance of the Northwestern formation; hence it will be advisable to inquire what else might cause that thinning and disappearance. There is of course the possibility of a low-angle fault, later folded, and that certainly cannot be disproved, but, in view of the low angle of discordance, it seems to be less probable than an unconformity. Moreover, the relationships in the Rochford district support the theory of unconformity better than that of faulting, for it would scarcely be expected that many miles away(many scores of miles along the strike) the Flag Rock formation would again rest on the Ellison formation if the contact were a fault. A more serious question is raised by the nature of the Northwestern formation itself; rather, by the manner in which it has been defined. For convenience in field mapping, the contact between Ellison and Northwestern formations has been drawn at the top of the highest quartzite, so that all the quartzites remain in the Ellison formation. This usage, a change from previous work which made the separation at a somewhat lower horizon, has been of great help in field mapping and is warranted because the maps show lithologic, not chronologic, boundaries. It is possible that a gradual upward extension of quartzites would cause the Northwestern formation to disappear; but, if so, there should be a thickening of the Ellison formation as the Northwestern formation thinned and this is not found. On the northeast edge of the district, thinning of the Northwestern formation is not accompanied by thickening of the Ellison formation. On the south edge of the district, and also in places beyond the limits of the map to the west and northwest, the Ellison formation has not thickened although all the Northwestern formation is gone (Fig. 1).

A relationship like that shown by the Ellison, Northwestern, and Flag Rock formations may be, according to Marr (1929, p. 82), the result of processes of sedimentation in a gulflike tract of comparatively small size during a cycle of deposit (i.e., between two terrestrial phases). In the subsiding basin, coarser sediments will be succeeded by finer ones, because of retreat of the shore lines; but if the shore lines subsequently converge, due to emergence of the land or to silting, shallow-water deposits will overlap the open-water ones. The result would be a lenticular mass of fine sediments underlain and overlain by coarser ones. This relationship does in fact agree with that shown by the Ellison, Northwestern, and Flag Rock formations (Fig. 1). However, such a mechanism probably requires that the cycle should begin and end with terrestrial phases, and the lower series of coarse sediments would prob-

ably rest on an unconformity. But in the Lead sequence there is no evidence of an unconformity below the Ellison formation.

Although the Flag Rock formation probably rests on an erosional unconformity, certain features commonly noted at unconformities are not found. Nowhere in the Lead or Rochford districts is there any basal conglomerate. Quartzites occur, but they were probably originally chert beds; moreover, they are not confined to the base of the formation, and the lowermost beds in some places are fine-grained graphitic slates or phyllites. Some evidence of an earlier period of weathering at the top of the Northwestern formation might have supported the idea of an unconformity. It is not easy to say just what change to expect, whether a change of color due to oxidation of ferrous minerals, a silicification, or a change in metamorphic minerals due to a change in total composition; but in fact there is no discernible change whatever. Finally, there is no support from igneous activity, because the earliest intrusive rocks are later than both sedimentary formations.

*Grizzly formation.*—The Grizzly formation in the map area barely reaches the outermost limit of the garnet zone, and the detailed descriptions are taken from outcrops within that zone. The characteristic material, in fact almost the only material, is a fine-grained, gray to dark-gray sericitic phyllite, with no distinctive character and almost no bedding. A few outcrops are brownish gray, and a few in the upper part of the formation are almost black. No horizon markers have been found, and the structure is undecipherable.

Microscopic study of the Grizzly formation phyllite gives little additional information. Quartz and muscovite are the dominant minerals, quartz usually being more abundant. Biotite is not common, but some streaks of oxides of iron parallel to schistosity may indicate the presence of some biotite in the fresh rock. Where biotite does occur, it is coarser than the muscovite; and there is some coarse white mica in some sections, in addition to the sericitic variety. A little tourmaline is present in several sections. Graphite is always present in at least small amounts and is abundant in some sections, in places producing a faint banding. The dimensional orientation of the muscovite plates is very good; and the quartz also shows good orientation, both dimensional and optical.

About 3000 feet of Grizzly formation has been observed in the map area, not allowing for possible duplication by undisclosed folding. The formation extends to Deadwood and probably is somewhat thicker there. Reconnaissance mapping has not disclosed the top of the formation.

*Cambrian Deadwood formation.*—Of the Paleozoic sedimentary rocks, only the Upper Cambrian Deadwood formation has been studied by detailed and reconnaissance mapping in the Lead district. The Deadwood formation begins with a basal conglomerate of well-rounded pebbles and boulders of vein quartz and a few pebbles and boulders of quartzite in a matrix of light-brown quartzite. Pebbles of schist or phyllite occur but are very rare. The pebbles and boulders range from an inch to 2 feet in diameter, and in a few places there are large angular boulders of quartzite. The conglomerate ranges from a single line of pebbles to about 50 feet thick; but in few places is it more than 10 feet thick. It merges with the overlying light-brown quartzite by the disappearance of the pebbles. Interbedding of conglomerate and quart-

zite occurs in the areas where the conglomerate is thickest, and in those places the quartzites have ripple marks, cross bedding, and other primary structural features; but in most of the district no bedding is visible. The quartzite ranges from 1 foot to, locally, 20 feet thick; but it exceeds 10 feet in few places. Over the quartzite is a brown impure dolomite. Where not too much altered (the dolomite has been extensively altered by hydrothermal solutions) the rock shows brown dolomite rhombs about 2 millimeters long. This member ranges from 2 feet to 15 feet thick, but thicknesses over 6 feet are rare in the Lead district. Thin partings of green shale are abundant in the upper part of the dolomite. The rest of the formation in the Lead district is thin-bedded green and gray glauconitic shale, with some narrow partings of limestone and brown dolomite and a few beds of intraformational limestone conglomerate. About 50 feet of shale is exposed in the map area below sills of Tertiary intrusive rocks, but in the Bald Mountain district, a few miles west, there is about 250 feet, which is overlain by more quartzite. Where the full thickness of the Deadwood formation is exposed, it ranges in thickness, in the northern Black Hills, from 300 to nearly 500 feet.

The unconformity at the base of the Deadwood formation is well exposed in the Lead district, in surface outcrops, open cuts, and some old mine workings. The erosion surface was very smooth; but there were a few broad swells and depressions and some minor projections. The thickness of the basal conglomerate is a good indicator of the irregularities. It thins or disappears over the ridges and swells, and it thickens to 50 feet in some depressions. The swells and minor projections are caused by the more resistant pre-Cambrian rocks. Quartzites, either in single wide beds or in zones, usually stand above the general level of the erosion surface and cut out the basal conglomerate, and the Homestake formation, being resistant to erosion, does the same. Not only the basal conglomerate, but also the Cambrian quartzite and dolomite are thinned or cut out over the projecting knobs. Beds of Ellison quartzite project above the Cambrian conglomerate, quartzite, and dolomite, so that shale rests directly on the pre-Cambrian quartzite. Whatever effects of weathering and oxidation there may have been cannot in most cases be distinguished from the recent ones, except the alteration which produced the Iron Dike in the Flag Rock formation. Where the unconformity has been seen in mine workings, effects of mineralization have destroyed any earlier effects of weathering.

*Gravel deposits.*—In several areas in the Lead district, no rock outcrops occur and the soil is thickly studded with large and small partly rounded boulders, mainly of Tertiary porphyry with a little Cambrian and pre-Cambrian quartzite and vein quartz. Some of this material has travelled a considerable distance from the nearest occurrence in place, and some has moved in directions different from those of the present drainage. In these areas the topography is very gentle, in marked contrast to that of most of the district. In places these gravel deposits contain clay beds several feet thick, and Darton (Darton and Paige, 1925, p. 2, 15) reports the presence of Oligocene fossils in the clays of these deposits. The most extensive area of gravels in the Lead district underlies the town of Lead and extends through the divides northwest, west, and southeast of the town and down Gold Run to the east. The maximum range of elevation observed in this area is 450 feet, and the maximum measured

thickness of gravel is about 90 feet. Other somewhat smaller areas occur in all parts of the district. In the Lead district, these gravel deposits do not overlap areas of Tertiary porphyry sills and stocks, and there is usually an increase in the size of boulders as the Tertiary igneous rocks are approached, from which it seems probable that the gravel deposits resulted from the destruction of mountain masses carved mainly from the Tertiary igneous rocks. The drainage pattern of the mature uplands between the youthful stream valleys is still nearly the pattern on which these gravels moved and accumulated in Oligocene time; but the gravels have been partly removed from the mature topography and completely removed from the youthful topography. The Tertiary gravels have probably been redeposited from time to time and in places grade into similar deposits of Quaternary age.

## IGNEOUS ROCKS

Two groups of igneous rocks occur in the Lead district; namely, the pre-Cambrian amphibolites, and the Tertiary porphyries. The amphibolites have been adequately described by Dodge (1942). They are believed to have been gabbroic rocks intruded near the end of the folding of the pre-Cambrian sedimentary rocks. They are widely distributed throughout the pre-Cambrian terrane of the entire Black Hills, in places more abundantly than at Lead. The distribution in the Lead area is shown on the map (Pl. 1). The amphibolites are mentioned in this paper on stratigraphy for two reasons: one, because typically they are long, narrow sills which might be mistaken for surface flows; and two, because metamorphism of the amphibolites in places has produced schists which are very difficult to distinguish from metamorphosed sedimentary rocks.

The group of Tertiary porphyries includes a number of closely related rocks, mainly of granitic and syenitic composition. They occur only in the northern Black Hills and are fairly abundant in the Lead district but are still more abundant northwest and west of Lead. The structural effects of the injection of the Tertiary intrusive rocks were very important, but the metamorphic effects were slight. The Tertiary intrusive rocks in the Lead district will be described in a separate paper.

## CORRELATIONS

We have made no serious attempts to correlate the Lead rocks with those of other parts of the Black Hills, except with the Rochford district, in the Black Hills about 20 miles south of Lead, where plane-table mapping of an area of about one-half square mile (a small part of the whole Rochford district) gave a repetition of the Lead sequence, from Poorman formation to Flag Rock formation, but without the Northwestern formation. The absence of the Northwestern is due to the unconformity at the base of the Flag Rock formation, which in part of the district also cuts out the Ellison formation. The Lead sequence is therefore repeated, except for the topmost member, the Grizzly formation, and there can be little doubt as to the correctness of the correlation. The Poorman formation is the same banded or dense, featureless ankeritic phyllite as at Lead. The Homestake formation in places is sideroplesite schist, in other places is cummingtonite schist; recrystallized chert pods are

always abundant. The Ellison formation has abundant dark-gray quartzite beds in soft-gray biotite phyllite. The black pyritic slates, streaked graphitic quartzites, and sideroplesite or cummingtonite members, so characteristic of the Flag Rock formation at Lead, have identical character in that formation at Rochford. Even the ore mineralization shows the same localization in the Homestake formation as it does at Lead. There is thus not only a repetition of distinctive formations or members, but a repetition of a sequence of distinctive formations, and the repetition of an unconformity. It is fortunate that the evidence for correlation is so positive, since the whole stratigraphic column is upside-down. The structure in the area mapped is that of a broad anticline plunging steeply to the south; but the youngest formation, the Flag Rock, is inside the arch of the fold. The probable mode of formation of this upside-down structure has been discussed by Harder (1934).

We know no other area in the Black Hills which can be surely correlated with the succession at Lead. Runner (1934) suggested that the Lead succession is repeated at Nemo, about 20 miles southeast of Lead, but the correlation is based on only three formations: a white limestone which is believed to correspond to the ankeritic phyllite of the Poorman formation; a quartz-hematite formation which is believed to correspond to the sideroplesite and chert of the Homestake formation; and very thick quartzites which are believed to correspond to the Ellison formation. But these are rock types not uncommon in all pre-Cambrian rocks of the sedimentary (Algonkian) type. While there is no evidence against the correlation between Lead and Nemo, there is only slight evidence for it. It seems probable that the Lead rocks recur in other areas in the Black Hills, but none is known. The most reasonable basis for correlation would be that already used at Rochford,—repetition of all or nearly all the sequence and the unconformity.

Pre-Cambrian rocks of sedimentary (Algonkian) type occur southwest of the Black Hills in the Hartville uplift in Wyoming, and northeast of the Black Hills in Minnesota. In the Hartville uplift, at the Sunrise property of the Colorado Fuel and Iron Company near Guernsey, Wyoming, the succession is white dolomite, an iron carbonate formation (which on alteration forms the iron ore), phyllites, and finally pyritic black slates and graphitic quartzites, locally altered to material like the Iron Dike at Lead[1]. The rocks are cut by pre-Cambrian intrusive rocks, now amphibolites. The sequence certainly resembles that at Lead, but as at Nemo, the rock types are not uncommon, at least in pre-Cambrian sedimentary rocks. Possibility of correlation with the Lake Superior region may come when a considerably larger area of the Black Hills has been mapped, but at this time any attempt at correlation could be only a guess.

## METAMORPHISM

### GENERAL

Harker (1932, p. 188) has stated that progressive or zonal metamorphism in a region can be studied and the relations of the zones established before the stratigraphy and

[1] The senior author is indebted to the officials of the Colorado Fuel and Iron Company and especially to Dr. J. B. Stone, former geologist for that company, for opportunity to visit the Sunrise property and study the geology.

tectonics are known, and he cites the study of metamorphism in the Scottish Highlands as an example. The study of the metamorphism in the Lead district gives no reason to question seriously this thesis although some qualifications will be given. On the other hand, progressive metamorphism, or an erratic distribution of metamorphism, may so hamper stratigraphic studies that it becomes impossible to determine the stratigraphy before the metamorphic changes are known. To some degree, this seems to be true in the Lead district. Although the zonal pattern of metamorphism can be seen in the progressive changes in each of the argillaceous rocks, the description of the stratigraphy of the district is confusing unless account is taken of the progressive metamorphism. All earlier discussion of the stratigraphy of the Lead area has suffered from this confusion, and the same is probably true of stratigraphic studies in some other pre-Cambrian areas.

In general, progressive metamorphism is best shown in sedimentary rocks of argillaceous composition. Rocks of this composition occur abundantly in all the sedimentary pre-Cambrian formations at Lead except the Homestake formation. The zones which have been distinguished in the argillaceous rocks of the Lead district are: a, biotite zone; b, garnet zone; and c, staurolite zone. The staurolite zone covers only a small part of the map area. The limits and distribution of the metamorphic zones are shown in the map (Pl. 1).

### BIOTITE ZONE IN THE ARGILLACEOUS ROCKS

The outermost zone is called the biotite zone, in accordance with the nomenclature of those writers who have studied the classic area of the Scottish Highlands, and the least metamorphosed rocks in the Lead district are very similar to the rocks of the biotite zone in the Scottish Highlands. The carbonate-free phyllites (Pl. 2, figs. 1, 2) consist essentially of quartz, muscovite or biotite (in places both micas), and minor amounts of graphite, tourmaline, sphene, and iron oxides. The biotite is nearly colorless or pale brown, in a few places pale green. Of many index determinations of biotite made by Dodge (1935, p. 44), most values for $\gamma$ fall between 1.625 and 1.652 $\pm$.004, which probably points to a composition between the iron-rich siderophyllite and the magnesium-rich eastonite (Winchell, 1927, p. 366–370). Muscovite tends to occur most abundantly in quartz-rich phyllites and biotite in those in which quartz is subordinate. Chlorite when present can be attributed to hydrothermal metasomatism, or in a few places possibly to retrograde metamorphism. A single thin section from several scores of thin sections shows what may be primary chlorite of an undetermined variety, partly replaced by biotite. All mineral grains are small, few mica plates exceeding 0.2 mm. in maximum diameter and few grains of quartz exceeding 0.1 mm. The detailed descriptions given in the chapter on stratigraphy apply mainly to the rocks of the biotite zone.

### GARNET ZONE IN THE ARGILLACEOUS ROCKS

The outcrop of the garnet zone in the Lead district is about 2 miles wide. The garnet isograd, marking the outer limit of the zone (Pl. 1), is drawn to include the farthest points at which garnets have been noted in the field mapping. The garnets occur very erratically in the outermost part of the zone, even in rocks of the most

favorable composition, some biotite-rich phyllites of the Poorman formation. The first garnets to appear are tiny lumps the size of a pin head; under the microscope they are seen to be about 0.5 millimeter in diameter. Some of these smaller grains are altered to sericite, quartz, and iron oxide, probably by retrogressive metamorphism facilitated by shearing during refolding of the rocks; and in a few places remnants of what may have been garnets can be found for short distances beyond the garnet isograd. The limits of these doubtful garnets are shown by a dotted line in Plate 1. No garnets have been found in argillaceous rocks by use of the microscope that were not visible to the eye; in fact, remnants of altered garnets are more readily detected in hand specimen than under the microscope. Systematic sampling and microscopic study probably would not extend the limits of the garnet zone.

A few hundred feet northeast of the garnet isograd, garnets become more abundant and appreciably larger. With the increase in the number and size of the garnets, there is an increase in grain size of all minerals and a disturbance of the rock cleavage. The presence of abundant garnets gives a wavy or lumpy appearance to cleavage surfaces, because micas wrap around the garnets. Colors change very little; but banding, by segregation of minerals along bedding planes, is much more common in the garnet zone than in the biotite zone. In the central portion of the garnet zone, the typical rock is garnet-mica schist (Pl. 2, figs. 3, 4), and only some light-colored rocks, consisting essentially of quartz and a little muscovite, have no garnets. In the inner part of the garnet zone, garnets become much more abundant and larger, and the whole rock becomes coarse grained, with micas up to 1.0 millimeter in maximum dimension. Here very few outcrops show no garnets, but certain garnet-free quartzose phyllites of the Ellison, Flag Rock, and Grizzly formations persist in small amounts throughout the garnet zone. There is probably a slight increase in the proportion of biotite to muscovite from the outer to the inner parts of the garnet zone. There is, however, no progressive increase in optical indices or depth of color of biotite with increasing grade of metamorphism.

On the basis of what is known of the composition of garnets in crystalline schists, it seems most probable that the garnet of these rocks is close to almandite. Gustafson (1930, p. 52) found indices of about 1.81 ±.01, indicating probably almandite plus a little pyrope and spessartite. Index determinations by Dodge (1935, p. 62) ranged slightly lower, between 1.80 and 1.81 ±.01, but still probably within the range of almandite.

Throughout most of the garnet zone the garnets carry abundant tiny inclusions of quartz, usually in narrow straight lines of grains, more rarely in symmetrical groupings of lines repeated by twinning of the garnet. In a few places there are tiny muscovite plates with the quartz inclusions. The mode of origin of the inclusions in garnets, and their significance in relation to folding, will be discussed in a paper on structure. Some of the garnets of the innermost part of the zone have narrow borders nearly or wholly free from inclusions. Associated with these are small garnets apparently free from inclusions; but these may be truncated corners of larger grains whose centers would still show abundant inclusions. In a few places, garnets of the inner part of the zone have grown into the narrow border of coarsely recrystallized quartz or into the micas which wrap around the garnets.

The mode of origin and early stages in the crystallization of the garnets of the phyl-

lites have not been observed with certainty. Ragged anhedral grains of garnet are not uncommon, and the first small grains usually have a rounded or subhedral outline. It may be that ragged grains indicative of early stages of growth of garnets have been destroyed by the retrogressive alterations already mentioned. On the other hand, the garnets may have started growth in rounded or subhedral grains. It is believed by most students of progressive metamorphism that almandite garnet of the garnet zone forms by combination of biotite and some remaining chlorite (Tilley, 1926; Phillips, 1930), and Harker has cited examples of incipient crystallization of garnet in a pattern apparently inherited from chlorite (Harker, 1932, p. 192). Such an explanation does not, however, accord with the observations in the Lead district, where garnets form in rocks which contain only quartz, micas, and iron oxides, without chlorite. The key to the origin of the garnets in this region is probably given by the retrogressive change, whereby garnets alter to an aggregate of quartz, sericite, iron oxide, and rarely biotite, but apparently not to chlorite. The same minerals probably combine to form almandite garnet with advancing metamorphism, thus:

$$\begin{array}{llllll} \textit{2 muscovite} & + \textit{3 quartz} + \textit{3 magnetite} = & \textit{3 almandite} + & \textit{solutions} \\ 2(OH)_2KAl_2Si_3O_{10} + & 3SiO_2 + & 3Fe_3O_4 & = 3Fe_3Al_2Si_3O_{12} + K_2O + H_2O + 3\,O \end{array}$$

If biotite enters the reaction, it probably produces some pyrope molecule. Some potash must be removed, and its ultimate disposition is not known; it may move out into the biotite zone and convert chlorite into muscovite and biotite. The early stages of crystallization of garnet in the quartzites also support this theory of origin of the garnets. Ambrose (1936) has found in Manitoba that almandite forms in rocks containing no chlorite, and he suggests that epidote, biotite, and magnetite unite to form almandite.

The origin of the widely distributed tourmaline of the Lead phyllites is not altogether clear. The common mode of occurrence is in short euhedral prisms which have a central core differing in color from the rest of the crystal. Most of these lie in the plane of schistosity; but there are always a few prisms transverse to the schistosity. The late formation, later than the development of garnets and later than all folding, would seem to point to a hydrothermal origin. The very erratic distribution, up to perhaps 2 per cent in some specimens but none in others, appears to support this origin; furthermore, there is some indication that tourmaline is more abundant in zones of greater shearing or refolding, although it is not more abundant in any particular metamorphic zone. On the other hand, in the coarser schists of the inner part of the garnet zone, where mineral particles are several times larger than in the biotite zone, the tourmaline prisms have also increased several fold. Furthermore, tourmaline of undoubted clastic origin, some in process of recrystallization, is found in Ellison quartzites in the garnet zone. If the tourmaline of the Lead rocks is of clastic origin, it has recrystallized at a late stage, since the refolding of the rocks, and it must have a recrystallizing range about equivalent to that of muscovite, which has also recrystallized in considerable amount since the refolding.

A surprising feature of the garnet zone is the appearance, about half way through the zone, of a white chlorite in coarse euhedral plates, increasing in size and abundance

throughout the inner part of the garnet zone but probably disappearing in the adjacent staurolite zone. The mineral can be seen in some hand specimens, where it resembles muscovite. Under the microscope it shows large plates, at random or oriented at a small angle to the schistosity (Pl. 2, fig. 5). Plates 1.0 mm. in maximum dimension occur in schists in which the next coarsest mica, usually biotite, is only half as large. In the coarsest schists, the chlorite attains 2.5 mm. in maximum dimension. The mineral is especially common as random grains associated with the areas of coarsely recrystallized quartz which fill the "dead ends" alongside garnet grains (Pl. 2, fig. 6). Long, tabular outlines with ragged ends are most common, and inclusions of quartz are seldom lacking. The shape, nature of inclusions, and random orientation closely resemble the usual mode of occurrence of chloritoid. The mineral in thin section is pale green, very faintly pleochroic, Z = colorless, X = Y = pale green, $\beta = 1.610 \pm .003$, birefringence almost zero to .006 (estimated), maximum interference color usually a dirty olive-green, rarely dull gray, 2V almost zero to 15° (estimated), optically positive. These data fit the chlorite mineral prochlorite, relatively high in magnesia and alumina, relatively low in iron (Winchell, 1927, p. 376; 1936, p. 649). This chlorite has been found in about half the schist sections from the inner part of the garnet zone. It seems to be dependent to some extent on suitable rock composition; but it is clearly also dependent on physico-chemical conditions, and its appearance indicates a certain grade of metamorphism. Possibly the prochlorite represents magnesia set free by replacement of biotite by garnet in excess of the amount of magnesia which can be incorporated in garnet under conditions of regional metamorphism. Another possible origin which may explain some of the prochlorite is as a product of sericitization of biotite in zones of cross folding. If desired, the prochlorite could be used to mark another isograd and so to restrict the garnet zone, by inserting a prochlorite zone between it and the staurolite zone, but this has not been done because the prochlorite is not readily recognized in the field or in hand specimens. The approximate location of the outer limit of prochlorite is shown by a dotted line on Plate 2. Chapman (1939) has noted an almost identical occurrence of prochlorite in New Hampshire, and a somewhat analogous occurrence of porphyroblasts of penninite has been described by Pilgrim and West (1928) in schists of a garnet zone in India.

### STAUROLITE ZONE IN THE ARGILLACEOUS ROCKS

The staurolite zone is characterized by the fairly abrupt appearance of coarse porphyroblasts of staurolite. As staurolite is dependent to a considerable extent on total composition of the rock, it has a somewhat erratic distribution; but it can be found by the microscope in some specimens which fail to show it in macroscopic sizes. Except for the addition of staurolite and a notable increase in grain size, the rocks of the staurolite zone do not differ in appearance from those of the adjacent inner part of the garnet zone. Only the Flag Rock formation enters the staurolite zone on the map area, and the rocks are moderately coarse-grained mica-garnet schists, in which staurolite occurs in crystals up to 4 centimeters long. Prismatic shapes are the rule, and there are a few oblique-angled twins but no cruciform ones. The mineral is

somewhat more common in light-colored schists with moderate amounts of small garnets than in the more highly garnetiferous schists. There is no indication, however, that garnets decrease in amount or size in the staurolite zone; in fact, few outcrops fail to show garnets in large amount.

The characteristic minerals of the staurolite schists are quartz, biotite, muscovite, garnet, staurolite, and small amounts of graphite and oxides of iron. Two sections had a few small random plates of a chlorite mineral, probably prochlorite. One section was unique in that it consisted almost wholly of mica and a little garnet and staurolite, with only one or two quartz grains; and the biotite had been converted to sericite, apparently as a result of shearing. Quartz and biotite usually form a groundmass for porphyroblasts of muscovite, garnet, and staurolite; muscovite in random plates up to 2.5 millimeters in maximum dimension, garnet in subhedral grains up to 5 millimeters in diameter, and staurolite in ragged or euhedral crystals up to several centimeters long (Pl. 2, figs. 7, 8). There is no progressive increase in optical indices or depth of color of biotite. Although biotite from different parts of the district has a considerable range in optical indices (Dodge (1935) recorded $\beta$ from 1.596 to 1.694), some of the low indices come from the staurolite zone and some high ones come from the biotite zone. Both garnet and staurolite show rotation in some sections, but coarse muscovite has crystallized since all movement ceased. Most crystals of staurolite, whether ragged or euhedral, contain abundant inclusions of quartz, and a few have inclusions of biotite.

Probably staurolite forms at the expense of biotite and muscovite, with the removal of some potash and the creation of some quartz as inclusions in the staurolite.

### PROGRESSIVE METAMORPHISM IN QUARTZITES

Rocks which can be called quartzites occur in three formations in the Lead district: in the Ellison formation; in the Flag Rock formation, probably originally chert; and in the Homestake formation, in 1 inch seams and pods probably originally chert. In outcrop and hand specimen the only change that can be observed in the Ellison quartzites, as the rocks pass into the higher metamorphic zones, is the disappearance of clastic grains. Clear rounded grains of quartz up to about 2 millimeters in diameter can be observed in many outcrops in the biotite zone and the outer part of the garnet zone but have not been observed in the quartzites of higher metamorphic grade.

The outstanding mineralogic difference between the phyllites and the Ellison quartzites is of course the great predominance of quartz in the quartzites, and this difference has two consequences: in the quartzites, the mica is commonly muscovite, biotite being rare until the inner part of the garnet zone is reached; and the first garnets appear only in the inner part of the garnet zone and there only as ragged grains of incipient crystallization. The commonly advanced and apparently reasonable explanation for the delay in formation of garnet and failure to form well-defined units is that overabundance of the relatively inert quartz hinders the metamorphic changes and delays crystallization (Harker, 1932, p. 243). The preponderance of muscovite over biotite, to complete exclusion of biotite in the quartzites of lower grade zones,

probably is due to the same cause, because muscovite probably indicates somewhat lower metamorphic grade than biotite.

Garnet is found in the Ellison quartzites only within the inner part of the garnet zone, as fixed by argillaceous rocks. There it is abundant, although not visible in the hand specimens because of the small size and ragged shapes. The garnet grains clearly show an incipient stage of crystallization. In places tiny irregular grains are strung out in straight lines parallel to bedding and schistosity of the quartzite, and in other places grains like these have coalesced to form long, narrow, very ragged seams of garnet (Pl. 3, figs. 1, 2). In very few sections are there grains approaching rounded outlines, and in none are there euhedral or subhedral forms. Further evidence on the origin of garnet is given by these quartzite specimens. The lines of tiny grains of garnet which first appear in the quartzites may well mark lines of tiny crystals of muscovite; and in fact one section shows a ragged linear group of garnet grains which apparently have replaced most of a line of muscovite plates (Pl. 3, fig. 2). There is no evidence in these sections that chlorite is essential to the formation of the garnets.

Prochlorite appears in a few small random grains in the Ellison quartzites of the inner part of the garnet zone. Not enough material has been studied to determine if its formation has been delayed over that in the argillaceous rocks, but at any rate the prochlorite was not delayed as much as was garnet. The plates are much smaller than those in the adjacent argillaceous rocks but otherwise have the same characteristics.

The origin of the tourmaline of the Ellison quartzites is fairly clear, because undoubted clastic grains are common, and there are many grains which have partly recrystallized. The clastic tourmaline grains were naturally somewhat coarser in the silts and sands which made the quartzites than they were in the finer grained sediments, and on account of this greater size the original clastic outlines are less easily obliterated by recrystallization. The original grains are browns and dark blues, whereas the recrystallized material, in optical continuity with the old, is pale yellow to light greenish blue.

Reference has been made to the disappearance of visible clastic grains in the outcrops of Ellison quartzite; and the same effect is seen under the microscope, although the clastic outlines, marked by borders of graphite and iron oxide (Pl. 3, figs 3, 4) are visible in all but the highest metamorphic grades, finally disappearing in the inner part of the garnet zone. When this point is reached, the recrystallized quartz grains, gradually becoming coarser with the increase of metamorphic grade, have attained a diameter of several tenths of a millimeter, and quartz has begun to segregate into narrow veinlike seams. Graphite and flakes of iron oxides segregate into small knots and groups. The polygonal pattern of quartz grains gives place to irregular grains with sutured contacts, very much like a hydrothermal pattern.

Previous descriptions mentioned the virtual absence of ordinary clastic impurities in the Flag Rock quartzites and in the tiny pods and bands of quartzite or recrystallized quartz in the Homestake formation. Clastic outlines of quartz grains have not been observed in the field or under the microscope, even in the lowest metamorphic grades, and this fact has been taken to indicate a chemical origin for these rocks. As might be expected from the composition, there is very little change in these quart-

zites in the higher metamorphic zones, except a gradual increase in grain size. In the Flag Rock quartzites of the inner part of the garnet zone, some quartz becomes segregated into bands of coarser grain size (up to 2 mm. in diameter), and these as well as the smaller grains show sutured rather than polygonal boundaries. In a few sections garnet occurs in ragged grains like those in the adjacent Ellison quartzites. At about the same point, there is a rather sudden appearance of cummingtonite and green hornblende. The cummingtonite forms in small stubby crystals to about 0.5 mm. long not uncommonly in crudely radial groups. Locally it makes up as much as 25 per cent of the rock. With it is a little green hornblende, in part replacing the cummingtonite and in part as small ragged or subhedral grains. There is also a very little dark-brown biotite and tourmaline, and the chlorite mineral daphnite occurs in small sharp plates in a few sections. It is possible that cummingtonite and hornblende here indicate progressive metamorphism, but it is more probable that they have been introduced hydrothermally, since these quartzites in lower grades have not carried impurities of the composition required for the formation of cummingtonite and hornblende. Cummingtonite is indeed abundant in the adjacent Homestake formation and in beds in the Flag Rock formation, where it has formed by regional metamorphism of sideroplesite, garnet, and biotite; and the possibility cannot be dismissed that the cummingtonite of the quartzites originated from grains of impurities which have not been detected in the Flag Rock quartzites of lower metamorphic zones. The conversion of cummingtonite to hornblende furthermore is probably in the direction of advancing metamorphism; but it seems improbable that the change would take place sooner in the quartzites than in the adjacent cummingtonite schist, and that shows no hornblendization.

### PROGRESSIVE METAMORPHISM OF LIME BEARING IRON CARBONATE ROCKS

The impure ankeritic phyllites of the Poorman formation in the biotite zone have been described, and the progressive metamorphism of these phyllites deserves some mention. In the lower grades of metamorphism, the rocks consist of ankerite, some muscovite and graphite, usually some quartz, and in places a little phlogopite; in a few specimens light-brown or greenish-brown biotite occurs instead of the much more common muscovite. In the ankeritic phyllites, there is no association of muscovite with greater abundance of quartz, such as was noted in the carbonate-free phyllites. No advance of metamorphism is perceptible until the inner part of the garnet zone is reached, where a few small garnets occur in some outcrops, and both muscovite and brown biotite occur. Quartz shows some tendency to segregate into distinct bands. The carbonate grains show no noticeable increase in average grain size. In one outcrop in the inner part of the garnet zone there is a small amount of a light-brown amphibole in the ankeritic phyllite. The carbonate by analysis has about 45 per cent $CaFeC_2O_6$, 55 per cent $CaMgC_2O_6$. The amphibole has $\beta = 1.662$ and can be either cummingtonite (somewhat low in iron) or hornblende, perhaps pargasite. In view of the abundance of lime in the carbonate, it seems probable that the amphibole is hornblende. Surprisingly, there are no other mineralogic changes in other outcrops of ankeritic phyllites in the garnet zone, although these impure phyllites contain the necessary constituents for a variety of metamorphic minerals It would seem that

ankerite with a moderately low iron content (here approximately 15 per cent $CaFeC_2O_6$) is somewhat more stable under advancing metamorphism than is sideroplesite or ankerite with somewhat higher iron content.

### PROGRESSIVE METAMORPHISM OF LIME-FREE IRON CARBONATE ROCKS

Because of its bearing on the nature of the ore mineralization, the metamorphism of the iron-magnesium carbonate (sideroplesite) schist of the Homestake formation has been the subject of considerable study, the most important work being that by Gustafson (1930; 1933). Although the Homestake formation contains most of the sideroplesite schist, narrow lenticular beds also occur in the Flag Rock, Poorman, and Ellison formations. The progressive metamorphism is identical in this material in all formations. Three mineralogic changes take place: cummingtonite forms at the expense of sideroplesite, about on the garnet isograd of the zones established by argillaceous rocks; garnet forms at the expense of biotite and quartz, at a somewhat higher grade; and more cummingtonite forms at the expense of quartz, garnet, and biotite, at still higher grade.

The sideroplesite schists are composed of sideroplesite, quartz, biotite, and usually a little chlorite (daphnite) and graphite. In a few specimens, sericitic white mica occurs instead of biotite. Sideroplesite commonly occurs in a mosaic of tiny polygonal grains, few exceeding 0.1 mm. in diameter (Pl. 3, fig. 5). In a few specimens the carbonate grains are distinctly flattened parallel to bedding, and in others they occur in scattered porphyroblasts up to 0.5 mm. in diameter. In a number of places, most of them mineralized areas, the carbonate grains are much coarser, reaching 5 mm. in diameter, and unbroken lines of graphite inclusions running through them probably indicate that they have grown by recrystallization. Most large grains are sheared and fractured. Some of the daphnite may be primary, but most is hydrothermal.

In spite of somewhat erratic distribution in a part of the Lead district, cummingtonite makes a definite zone in the lime-free iron carbonate rocks, and the outer limits of this zone can be fixed almost as exactly as can that of garnet in the argillaceous rocks. As well as can be determined from the distribution of the lime-free iron carbonate rocks, the cummingtonite isograd coincides exactly with the garnet isograd of the argillaceous rocks. This apparently fortuitous coincidence may be of genetic significance. If we may suppose that a garnet of somewhat different composition would normally appear in the lime-free iron carbonate rocks under the metamorphic conditions whereby almandite garnet forms in the argillaceous rocks, this coincidence of garnet and cummingtonite isograds could result if this garnet immediately reacted with quartz and biotite to form cummingtonite. That such a reaction can occur is indicated in a subsequent paragraph describing the increasing development of cummingtonite in the inner part of the garnet zone where sideroplesite has disappeared.

No cummingtonite has been found southwest of the garnet isograd of Plate 1. In the first few hundred feet northeast of that isograd, cummingtonite replaces sideroplesite schist in small scattered blades or as sparsely scattered large blades, usually in a crudely radial arrangement (Pl. 3, figs 5, 6). Here the distribution of cummingtonite is erratic; small areas of cummingtonite schist occur alongside areas of sideroplesite schist with no cummingtonite. It seems reasonable to suppose that differ-

ences in permeability have something to do with this erratic distribution, because in the conversion of sideroplesite to cummingtonite the carbonate radical must be eliminated.

*7 sideroplesite + 8 quartz + water = cummingtonite + 7 carbon dioxide*

$$7\,(Fe,Mg)CO_3 + 8\,SiO_2 + H_2O = (OH)_2(Fe,Mg)_7(Si_4O_{11})_2 + 7\,CO_2$$

Gustafson (1933, p. 144) believes that the quartz is primary quartz of the sideroplesite schist. This may be true for the whole formation, but the freedom of movement during the metamorphism must have been somewhat greater than is ordinarily supposed, because quartz-free carbonate bands an inch or two wide are as readily converted to cummingtonite as are those in which both sideroplesite and quartz occur. Beyond the first few hundred feet from the garnet isograd, sideroplesite is practically absent, and the replacement of sideroplesite by cummingtonite is complete.

Physical and chemical data for the Lead cummingtonite are given by Wayland (1936) and by Gustafson (1930).

Farther northeast, about the middle of the garnet zone of the argillaceous rocks, garnet appears in the cummingtonite schist. The composition of this garnet has not been determined. It is found in most specimens but is not as abundant as in the adjacent mica phyllites. In some sections the garnet shows tiny irregular or platy forms which probably are inherited from muscovite or biotite, and presumably it has the same origin as ascribed to almandite in the argillaceous rocks. Biotite occurs both as small plates with the polygonal quartz grains and as coarse stubby grains, both brown and green, which are either recrystallized or introduced by hydrothermal solutions. Quartz has decreased in some sections but is fairly abundant in others. Minerals which are surely hydrothermal include daphnite, several sulphides, and a carbonate, probably ankerite. In some of the lenticular beds of cummingtonite schist of the Flag Rock formation, cummingtonite, biotite, and garnet seem to have been recrystallized during a hydrothermal stage of the metamorphism.

A still higher grade of metamorphism is shown by the Homestake specimens from the inner part of the garnet zone. Here cummingtonite becomes very abundant and is almost the only constituent in some specimens. Accompanying the increase of cummingtonite, quartz decreases and in fact has disappeared from some specimens—even the inclusions of quartz in garnets. Graphite is conspicuous locally, perhaps because it is largely concentrated in the small amount of residual quartz. A considerable amount of a plagioclase feldspar, probably andesine, was noted in one specimen, but this may have been introduced hydrothermally. Small amounts of a late-forming dark-brown biotite occur in most specimens. Hydrothermal minerals noted in a few places are daphnite, a carbonate, sulphides, and perhaps some of the biotite. All specimens show garnet in process of replacement by cummingtonite (Pl. 3, figs. 7, 8), and quartz is being replaced by cummingtonite in all sections in which it still remains. Clearly the presence of sideroplesite is not essential to the formation of cummingtonite, because sideroplesite disappeared in the outer part of the garnet zone. A reaction may take place analogous to the following.

*quartz + biotite + almandite = cummingtonite + solutions*

$$SiO_2 + H_4K_2(Fe,Mg)_6Al_4Si_6O_{24} + Fe_3Al_2Si_3O_{12} = (Mg,Fe)SiO_3 + H_2O,\ K_2O,\ Al_2O_3$$

To the extent that the garnet contains the pyrope molecule, less biotite will be demanded to furnish magnesium for the cummingtonite. When rocks in which sideroplesite has already been completely replaced show continued growth of cummingtonite at the expense of quartz, biotite, and almandite, there can be no reasonable doubt that potash and alumina have been removed. If a reaction of this sort did not take place, then the increasing amount of cummingtonite must be explained as a product of hydrothermal solutions. But the increasing cummingtonite is merely a part of the zonal distribution of cummingtonite in lime-free iron carbonate rocks, just as the increasing amount of garnet is part of the zonal distribution of garnet in argillaceous rocks. It seems doubtful that hydrothermal solutions would have made a highly selective but complete replacement of the cummingtonite-bearing schists while preserving the pattern of zonal metamorphism.

A suite of specimens of cummingtonite schist collected at intervals from inner to outer parts of the garnet zone was investigated by Wayland (1936), who found that the cummingtonite of different specimens differed somewhat in ferrous iron content and in optical properties, but that the differences showed no relations to the areal distribution.

### PROGRESSIVE METAMORPHISM OF AMPHIBOLITES

Dodge (1935; 1942) has shown that the amphibolites of the Lead district have been formed by metamorphism of intrusive rocks probably originally gabbroic. Rocks of the same general type are widely distributed in the pre-Cambrian terrane of the Black Hills and have been mapped and described by Paige (Darton and Paige, 1925). In the Lead district the amphibolites are for the most part confined to a belt which occupies the eastern half of the district and extends north and south beyond the limits of mapping, crossing the metamorphic zones at a small angle (Pl. 1). Most of the amphibolites are composed principally of hornblende and fine-grained secondary plagioclase, with subordinate amounts of biotite, apatite, and ilmenite, usually some quartz, and usually some alteration minerals, mainly calcite, chlorite, and sericite. A type with coarse lath-shaped feldspars more nearly represents the primary rock, but it too has suffered some metamorphism. Another type has abundant garnets.

At the time of Dodge's field work the progressive metamorphism of the Lead rocks had been recognized, but no attempt had been made to map the isograds or isothermal lines. The validity of the zonal theory was verified by the microscopic study of the amphibolites, and the following progressive features were noted (Dodge, 1935; 1942):

a. In specimens from the northeastern part of the district, hornblende changes to higher indices ($\beta$ increases from 1.670 to 1.680) and higher intensity of color.

b. In specimens from the northeastern part of the district, feldspars are more calcic (in place of oligoclase, the feldspar is andesine).

c. The grain size shows a progressive increase toward the northeast.

d. Garnets occur in all amphibolite specimens from the northeastern part of the district but not elsewhere (except for one small area); they first appear in the inner part of the garnet zone of the argillaceous rocks.

In addition to the amphibolites of the belt in the eastern part of the district, there

is, on the western edge of the district, a narrower belt containing rocks which on basis of mode of occurrence were mapped as amphibolites, but on basis of mineralogy, Dodge (1935, p. 252) called most of them not amphibolites but altered phases of the sedimentary schists. A reasonably smooth line, with general north trend, separates all the unquestioned amphibolites from the doubtful ones, with one or two exceptions. Most of the doubtful rocks consist almost wholly of chlorite, calcite, and quartz, with minor sphene, apatite, and iron oxides; a few rocks (called amphibolites by Dodge) consist of tremolite, talc (or sericite), serpentine (or chlorite), dolomite, and magnetite; and one group consists of chlorite, quartz, and sericite, plus some remnants of hornblende and feldspar. We are of the opinion that these rocks are not alteration products of sedimentary schists, but have been derived from gabbroic intrusive rocks, with a history partly in common with the normal amphibolites of the eastern belt. The differences between the normal and the doubtful types can be explained as due to progressive metamorphism of altered gabbroic rocks (metamorphosed during crystallization), or as due to later retrogressive metamorphism. Two facts argue against progressive metamorphism as an explanation: there is a small but definite angle between the isograds of the schists and the dividing line between normal and doubtful amphibolites; and there is fairly definite evidence that the minerals characteristic of the doubtful rocks replace those characteristic of the normal ones. We believe therefore that the doubtful rocks of the western belt are alteration products of the normal amphibolites, the alteration being a retrogressive one initiated by strong shearing and effected by moving solutions which have caused some changes in bulk composition.

## MINERAL FACIES

The principal mineralogic changes in the metamorphic rocks of the Lead district can be conveniently shown by the facies classification of Table 1. Minerals in parentheses, in process of replacement, indicate lack of equilibrium. This table shows the extent to which the appearance of new metamorphic minerals is delayed in some types of rocks, especially in quartz-rich rocks and in ankeritic phyllite. A proper understanding of this effect of delay is of some importance in stratigraphic studies. Harker's statement (1932, p. 188) that progressive or zonal metamorphism can be studied and the relations of the zones established before stratigraphy is known is probably true only if due allowance is made for such delayed appearances. The usual caution to establish the metamorphic zones in rocks of argillaceous composition should be supplemented by caution to avoid those argillaceous rocks moderately rich in quartz. It is clear that the sandstones and cherts of the Lead sequence lag far behind the shales in metamorphic changes and would be less suitable for metamorphic studies. What is not so clear except on close study is that light-colored phyllites moderately rich in quartz lag behind the more normal darker ones in which quartz is subordinate. This difference has been of considerable importance in the Lead district because the upper part of the Poorman formation and the phyllites of the Ellison formation can be distinguished in most of the mine area by presence or absence of garnets; the quartz-deficient Poorman formation is rich in biotite and has abundant garnets, but the quartz-rich Ellison phyllites are muscovite phyllites with

TABLE 1.—*Mineral facies of the metamorphic zones in pre-Cambrian rocks of the Lead district**

| Original Rock | Zones of Progressive Metamorphism | | | | |
|---|---|---|---|---|---|
| | Biotite Zone | Garnet Zone | | | Staurolite Zone |
| | | Outer | Middle | Inner | |
| Shale | Quartz<br>Muscovite<br>Biotite<br>minor constit. | Quartz<br>Muscovite<br>Biotite<br>Almandite | | Quartz<br>Muscovite<br>Biotite<br>Almandite<br>Prochlorite | Quartz<br>Biotite<br>Almandite<br>Muscovite<br>Staurolite<br>(Prochlorite) |
| Sandstone and Chert | Quartz<br>little Muscovite<br>minor constit. | | | Quartz<br>little Muscovite, Biotite,<br>Almandite, Prochlorrit e<br>(disappearance of clasti c grains) | |
| Impure iron-bearing Dolomite | Ankerite<br>Muscovite<br>Quartz<br>little Biotite, Phlogopite | | | Ankerite<br>Quartz<br>Biotite<br>Almandite<br>little Horn-blende<br>(Muscovite) | (?)<br>(absent) |
| Fe-Mg Carbonate Rock | Sideroplesite<br>Quartz<br>Biotite<br>Daphnite (?) | Cummingto-nite<br>Quartz<br>Biotite<br>(Sideroplesite) | Cummingto-nite<br>Quartz<br>Biotite<br>Almandite | Cummingto-nite<br>(Almandite)<br>(Biotite)<br>(Quartz) | (?)<br>(absent) |
| Intrusive Gabbro (etc.) | Hornblende ($\beta$ = 1.670)<br>Oligoclase (albite) etc. | | | Hornblende ($\beta$ = 1.680)<br>Andesine (oligoclase)<br>Garnet etc. | |

*Distinctive new minerals are underlined
Disappearing minerals are in parentheses

no garnets. Outside the mine area, however, this basis of distinction disappears; to the northeast both formations are biotite-garnet schists, to the southwest both are muscovite phyllites without garnets. Another example of the possible confusion caused by delayed appearance of metamorphic minerals is seen along City Creek (on the northeast edge of the map area, (Pl. 1)). The ankeritic phyllites of the Poorman formation are almost devoid of garnets, but the adjacent Ellison formation, the Northwestern formation, and the Flag Rock formation carry abundant garnets, and the latter also has staurolite. Continuing east (just beyond the limits of mapping), the quartz-rich phyllites of the Grizzly formation on outcrops show only a few

tiny garnets, and it appears that we are about to leave the garnet zone; only by microscopic study do we find an abundance of garnets.

### CHANGES OF COMPOSITION DURING METAMORPHISM

It has been indicated that some of the postulated chemical reactions, based on microscopic data and assumed mineral compositions, point to probable changes in bulk composition during metamorphism. In the absence of accurate sampling of known stratigraphic horizons, we can only guess at the amount of change, but the direction of change, if any exists, can be deduced from theoretical considerations. Of the five types of rock listed in Table 1, only three have mineralogic changes sufficiently distinctive to serve as a basis for discussion; these are the shales or phyllites, the sideroplesite-cummingtonite rocks, and the amphibolites. If we assumed that the argillaceous rocks of the Lead district were once chloritic slates and have been raised to the present metamorphic grades by an advance of isothermal surfaces, then we can assume the following reactions:

a. To produce the biotite of the biotite zone:

1. Chlorite and muscovite and iron oxides and silica form biotite.
2. Chlorite and potash form biotite.

The first reaction is that commonly quoted (Harker, 1932, p. 49) the second reaction is here held to be more probable.

b. To produce the almandite and prochlorite of the garnet zone:

3. Micas and quartz and magnetite form almandite and set free potash.
4. Micas and alumina and magnesia form prochlorite and set free potash.

c. To produce the staurolite of the staurolite zone:

5. Micas form staurolite and quartz and set free potash.

Thus in all reactions of zones above the biotite zone, potash is set free and probably removed from the rocks. On the other hand, it is possible that potash is utilized in the biotite zone, to make over chlorite into biotite. While there is nothing compelling in the argument, there is at least a suggestion of an outward migration of potash ahead of the advancing isothermal surfaces. Phillips and Hess (1936, p. 354, 355) found a somewhat similar removal of alkalies from metamorphic zones around serpentinites in Vermont and Quebec. On the other hand, Barth (1936) has listed analyses which are interpreted as showing that the alkali-alumina ratio increases with advance of metamorphism in the metamorphosed rocks of southeastern New York. Billings (1938), however, found no increase in the potash-alumina, soda-alumina, or alkali-alumina ratios with advance of metamorphism; if anything, there was a decrease in these ratios.

It is also possible to suggest the probable reactions in the progressive metamorphism of sideroplesite schist to cummingtonite, and here the mineralogy is somewhat simpler.

a. To produce quartz-cummingtonite-biotite schist (in the garnet zone):

1. Quartz and sideroplesite form cummingtonite and set free $CO_2$.

b. To produce quartz-cummingtonite-biotite-garnet schist (in the inner part of the garnet zone):

2. Quartz and mica and magnetite form almandite and set free potash.

c. To produce cummingtonite schist (in the staurolite zone):

3. Quartz and biotite and almandite form cummingtonite and set free potash and alumina.

There is once more a suggestion that potash is driven ahead of the advancing isothermal surfaces. If an outer zone consists of sideroplesite and chlorite, the chlorite by addition of potash might be converted to biotite.

The chemical changes involved in the mineralogical changes of the amphibolites are not well known, and no notable changes in bulk composition are indicated. In the alkalies, the changes in composition of hornblende and feldspar nearly balance. Some lime may be driven out of the rocks of the outer zone, but it could easily remain in the rocks as calcite. However, the alterations of amphibolites essentially of retrogressive and hydrothermal character call for changes in chemical composition.

It is worthy of note that all these reactions are in the direction of the elimination of lighter, more volatile constituents and the creation of denser minerals in place of lighter ones.

## PROBABLE SHAPES OF THE ISOTHERMAL SURFACES

The underground workings of the Homestake mine give some opportunity to estimate the shape of a small part of the garnet isothermal surface. The garnet isograd on the surface follows very close to the west edge of the open cut, and the limit of cummingtonite follows the same line. In mine levels about 1000 feet below the surface, a few porphyroblasts of garnet and cummingtonite are found for about 1000 feet west of the position of the line on the surface. There is some indication, therefore, that in this place the isothermal surface dips outward, to the west, probably at a moderately low angle, with the garnet zone underlying the biotite zone.

The relation of the metamorphic zones of the Lead area to the metamorphism of the rest of the pre-Cambrian rocks of the Black Hills is known only in a general way. West of Lead, in the small area of pre-Cambrian rocks between Lead and Spearfish Canyon, garnets have not been found and probably do not occur. That area probably all lies in the biotite zone. Southward from Lead many areas are free from garnets, but small areas of sparsely garnetiferous schists are widely distributed, and some prochlorite is found. At Rochford, 20 miles south of Lead, there are some small garnets and in the iron-rich carbonate rocks some cummingtonite, but there is also much chlorite which appears to be a primary constituent of the rocks. It is possible that part of the erratic distribution of garnets south of Lead is due to presence of rocks of unsuitable composition, but it seems more probable that the present erosion surface has been cut close to a flat-lying but somewhat irregular isothermal surface between the biotite zone and the underlying garnet zone.

Garnetiferous schists occur abundantly but erratically in the southern Black Hills in the vicinity of the intrusions of Harney granite. In the general region of Keystone there are some occurrences of staurolite schist which are closely confined to the vicinity of pegmatite dikes. Alongside a dike near the Keystone mine staurolite crystals are found to distances of 100 feet from the dike walls, garnets to distances of 300 feet, and coarse muscovite schist beyond that. In this case the isothermal surfaces must conform to the attitude of the dike and are nearly vertical. The oc-

currence at Keystone is part of a semicircular zone which extends from Hayward through Hill City and Oreville to Custer, about parallel to the border of the Harney Peak granite batholith (Lincoln and Cummings, 1937). Within this zone there is also a little andalusite and sillimanite.

### CAUSE OF THE PROGRESSIVE METAMORPHISM

Progressive metamorphism has been described in a number of places in the world, and different causes have been assigned by different students. For the progressive metamorphism of the Lead district, we believe that the most probable cause is the presence of undisclosed masses of intrusive igneous rock to the northeast of the district. There are two other causes, however, which have been proposed to explain relations in other regions, and they cannot be dismissed as impossible. Pressure and heat due to depth of burial have been suggested as causes of regional metamorphism. They should produce nearly horizontal zones which could subsequently be deformed and then exposed by erosion. The sporadic occurrence of garnetiferous schists throughout much of the Black Hills pre-Cambrian terrane does suggest nearly flat zones; but the width of zones in the Lead district is an argument against depth of burial as a main cause of the zones. The garnet zone has an outcrop width of only 2 miles and probably has an attitude somewhat flatter than 45 degrees; the thickness at right angles to the isothermal surfaces is therefore somewhat less than 1½ miles. The normal increment of pressure and temperature for this moderate depth would seem to be totally inadequate to account for the mineralogical changes which take place in the garnet zone.

Another cause of progressive metamorphism proposed for many regions is that of difference in intensity of rock deformation. Most of the recent writers on the metamorphism of the Scottish Highlands favor this cause (Bailey, 1923; Tilley, 1925; Elles and Tilley, 1930; Harker, 1932, p. 179, 185–189), although most early writers preferred thermal zoning due to igneous intrusions (Barrow, 1893; 1912; Hill, 1899; Phillips, 1930; Read, 1939.). In the Lead district, the dynamic history, though complex, is sufficiently well known to indicate definitely that there is no relation between intensity of rock deformation and the progressive metamorphism. The rocks have been subjected to a uniform and very intense compression, followed by shearing deformations probably in several distinct zones. There is no indication of any difference in the intensity of the first period of deformation, and no relation between the deformation of the second period and the zones of progressive metamorphism.

The conclusion that the progressive metamorphism of the Lead district is attributable to undisclosed masses of intrusive igneous rock is therefore adopted partly by elimination of other possible causes; but there is a little supporting evidence. The occurrence of narrow garnet and staurolite zones bordering a pegmatite dike near Keystone has been mentioned; the intrusion there is clearly the direct or indirect cause of the metamorphic zones, and by analogy the progressive metamorphism of the Lead district can be ascribed to a similar cause. The presence of pegmatite dikes on Whitewood Peak, 5 miles northeast of Lead, shows that igneous activity has occurred in that part of the district. Direct evidence of intrusive masses will not be obtained, because the pre-Cambrian rocks are covered by younger rocks just beyond

the northeast edge of the map area of Plate 1, except the small area on Whitewood Peak.

To believe that masses of intrusive igneous rocks have caused the progressive metamorphism of the Lead district is not to believe that the metamorphism is wholly thermal. The injection of a large mass of intrusive igneous rock must have several consequences in the invaded rocks, only one of which is an increase of temperature. It can scarcely be doubted that the invaded rocks must experience shearing stresses of varying intensity, and perhaps in addition an increase in hydrostatic pressure, and that hydrothermal solutions will pass through the invaded rocks, perhaps to great distances and in great amount. The progressive metamorphism in the rocks of the Lead district is believed to be a result of this complex chain of factors.

### RETROGRESSIVE METAMORPHISM

Retrogressive metamorphic effects are not conspicuous in the rocks of the Lead district, but in small amounts they are widely distributed. Probably the most common effect is the alteration of almandite. In many places the almandite grains are slightly altered to a very fine-grained aggregate of quartz, sericite, iron oxide, and in places a little biotite. In a few places the garnets are completely altered. A dashed line on Plate 1 shows the outer limit of what may be completely altered garnets which occur beyond the garnet isograd, and completely altered garnets occur in a few other small areas within the garnet zone. In a number of places, biotite seems to have been converted to sericite by shearing, and this can be classed doubtfully as retrogressive metamorphism.

## REFERENCES CITED


Ambrose, J. W. (1936) *Progressive kinetic metamorphism in the Missi Series near Flinflon, Manitoba*, Am. Jour. Sci., 5th ser., vol. 32, p. 257–286.

Bailey, E. B. (1923) *The metamorphism of the South-west Highlands*, Geol. Mag., vol. 60, p. 317–331.

Barrow, George (1893) *On an intrusion of muscovite-biotite gneiss in the southeastern Highlands of Scotland and its accompanying metamorphism*, Geol. Soc. London, Quart. Jour., vol. 49, p. 330–358.

——— (1912) *On the geology of the lower Dee-side and the Southern Highland Border*, Geol. Assoc. London, Pr., vol. 23, p. 274–290.

Barth, T. F. W. (1936) *Structural and petrologic studies in Dutchess County, New York: part II, Petrology and metamorphism of the Paleozoic rocks*, Geol. Soc. Am., Bull., vol. 47, p. 775–850.

Billings, M. P. (1938) *Introduction of potash during regional metamorphism in western New Hampshire*, Geol. Soc. Am., Bull., vol. 49, p. 289–302.

Bruce, E. L. (1945) *Pre-Cambrian iron formations*, Geol. Soc. Am., Bull., vol. 56, p. 589–602.

Chapman, C. A. (1939) *Geology of the Mascoma quadrangle*, Geol. Soc. Am., Bull., vol. 50, p. 127–180.

Darton, N. H., and Paige, Sidney (1925) *Description of the Central Black Hills*, U. S. Geol. Survey, Geol. Atlas, Folio no. 219.

Dodge, T. A. (1935) *The amphibolites of the Lead area, South Dakota*, Harvard Univ., doctorate thesis.

——— (1942) *Amphibolites of the Lead area, northern Black Hills, South Dakota*, Geol. Soc. Am. Bull., vol. 53, p. 561–584.

Elles, G. L., and Tilley, C. E. (1930) *Metamorphism in relation to structure in the Southern Highlands*, Royal Soc. Edinburgh, Tr., vol. 56, pt. III, p. 621–646.

Gruner, J. W. (1922) *The origin of the sedimentary iron formations: the Biwabik formation of the Mesabi Range*, Econ. Geol., vol. 17, p. 407–460.

Gustafson, J. K. (1930) *The Homestake gold-bearing formation*, Harvard Univ., doctorate thesis.
——— (1933) *Metamorphism and hydrothermal alteration of the Homestake gold-bearing formation*, Econ. Geol., vol. 28, p. 123–162.
Harder, J. O. (1934) *Geology of a pre-Cambrian area at Rochford and its relation to the regional structure of the northern Black Hills*, S. Dak. School Mines, Master's thesis.
Harker, Alfred (1932) *Metamorphism*, Methuen, London, 356 p.
Hill, J. B. (1899) *On the progressive metamorphism of some Dalradian sediments in the region of Loch Awe*, Geol. Soc. London, Quart. Jour., vol. 55, p. 470–493.
Hosted, J. O., and Wright, L. B. (1923) *Geology of the Homestake orebodies and the Lead area of South Dakota*, Eng. Min. Jour. Press, vol. 115, p. 793–799, 836–843.
Leith, C. K., Lund, R. J., and Leith, A. (1935) *Pre-Cambrian rocks of the Lake Superior region*, U. S. Geol. Survey, Prof. Paper 184, 34 p.
Lincoln, F. C., and Cummings, J. B. (1937) *Pegmatite mining in South Dakota*, S. D. State Planning Board, Brookings, 80 p.
Marr, J. E. (1929) *Deposition of the sedimentary rocks*, Cambridge Univ. Press, Cambridge, 239 p.
Marsden, R. W. (1933) *Accessory minerals of the Harney Peak granite*, Univ. Wis., Master's thesis.
McLaughlin, D. H. (1931) *The Homestake enterprise, ore genesis and structure*, Eng. Min. Jour., vol. 132, p. 640–645.
Paige, Sidney (1924) *Geology of the region around Lead, South Dakota*, U. S. Geol. Survey, Bull. 765, 56 p.
Pettijohn, F. J. (1943) *Archean sedimentation*, Geol. Soc. Am., Bull., vol. 54, p. 925–972.
Phillips, A. H., and Hess, H. H. (1936) *Metamorphic differentiation at contacts between serpentine and siliceous country rock*, Am. Mineral., vol. 21, p. 333–362.
Phillips, F. C. (1930) *Some mineralogic and chemical changes induced by progressive metamorphism in the Green Bed group of the Scottish Dalradian*, Mineral. Mag., vol. 22, p. 240–256.
Pilgrim, G. E., and West, W. D. (1928) *The structure and correlation of the Simla rocks*, Geol. Survey India, Mem., vol. 53.
Read, H. H. (1939) *Metamorphism and igneous action*, Nature, vol. 144, p. 729–731, 772–774.
Runner, J. J. (1934) *Pre-Cambrian geology of the Nemo district, Black Hills, South Dakota*, Am. Jour. Sci., 5th ser., vol. 28, p. 353–372.
Tilley, C. E. (1925) *A preliminary survey of the metamorphic zones in the Southern Highlands of Scotland*, Geol. Soc. London, Quart. Jour. vol. 81, p. 100–112.
——— (1926) *Some mineralogic transformations in crystalline schists*, Mineral. Mag., vol. 21.
Van Hise, C. R., and Leith, C. K. (1911) *The geology of the Lake Superior region*, U. S. Geol. Survey, Mon. 52, 641 p.
Wayland, R. G. (1936) *Cummingtonite from the Black Hills, South Dakota*, Am. Mineral. vol. 21 p. 607–610.
Winchell, A. N. (1927) *Elements of optical mineralogy, pt. II*, John Wiley and Sons, New York, 424 p.
——— (1936) *A third study of chlorite*, Am. Mineral. vol. 21, p. 642–651.

California Institute of Technology, Pasadena 4, Calif.; Homestake Mining Company, Lead, S. D.
Manuscript Received by the Secretary of the Society, August 15, 1947

## EXPLANATION OF PLATES 2 AND 3

## PLATE 2.—PHOTOMICROGRAPHS OF TYPICAL SECTIONS OF METAMORPHIC ROCKS FROM THE LEAD DISTRICT

ALL × 20

FIGURE 1. Mica schist from the biotite zone. Consists of quartz, fine-grained light-green biotite, coarser brown biotite, a little magnetite and graphite, and a few grains of tourmaline. The green biotite is well oriented, the quartz is fairly well oriented both dimensionally and optically, and the brown biotite is partly oriented and partly random. Plane polarized light.

FIGURE 2. Same field as a, but x-nicols.

FIGURE 3. Mica-garnet schist from the outer part of the garnet zone. Consists of biotite, muscovite, garnet, quartz, and a little magnetite and graphite. The micas are well oriented, and quartz is in good dimensional orientation, but the orientations are disturbed near the garnets. The garnets contain small inclusions of quartz and are replaced around the edges and along fractures by carbonate. Plane polarized light.

FIGURE 4. Mica-garnet schist from the inner part of the garnet zone. Consists of biotite, garnet, quartz, and a little tourmaline and iron oxide. There is some orientation of biotite and quartz, but the orientation is poor because of the abundance of garnets. The garnets contain abundant fairly large inclusions of quartz. Plane polarized light.

FIGURE 5. Prochlorite in mica-garnet schist from the inner part of the garnet zone. Consists of biotite, muscovite, quartz, prochlorite, and a little graphite and tourmaline. Elsewhere the section has small garnets. The micas and quartz are well oriented, and a large ragged crystal of prochlorite cuts the schistosity at a low angle in a faint zone of cross folding. Prochlorite replaces biotite and encloses quartz. The colorless areas alongside the prochlorite are quartz which has recrystallized since the growth of the prochlorite. Plane polarized light.

FIGURE 6. Prochlorite in mica-garnet schist from the inner part of the garnet zone. Consists of biotite, muscovite, quartz, garnet, prochlorite, and a little tourmaline, iron oxide, and graphite. The micas and quartz wrap around the garnet crystal, and the micas, quartz, and prochlorite have recrystallized in the "dead end" alongside the garnet crystal. Prochlorite is probably the last to form. The garnet contains small quartz inclusions, and it has iron oxide in fractures and around the edges of the grain. Plane polarized light.

FIGURE 7. Mica-garnet-staurolite schist from the staurolite zone. Consists of staurolite, garnet, biotite, quartz, and a little magnetite and graphite. There is no noticeable orientation of mica and quartz in this field. The centers of the garnets contain a few quartz inclusions. The staurolite crystals contain abundant quartz inclusions. Plane polarized light.

FIGURE 8. Mica-garnet-staurolite schist from the staurolite zone. Consists of staurolite, biotite quartz, and a little magnetite and graphite. Elsewhere the section contains garnet. Biotite and quartz show good dimensional orientation, but the orientation is disturbed near the staurolite crystal. The staurolite is full of small lines of quartz inclusions. Plane polarized light.

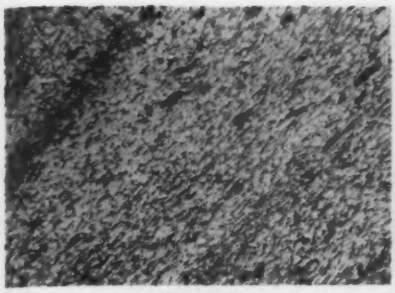


FIGURE 1

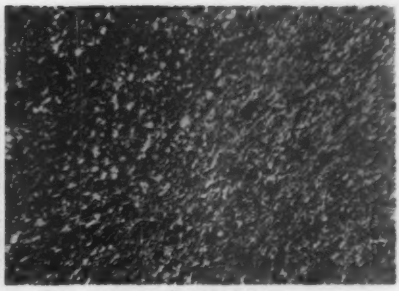


FIGURE 2

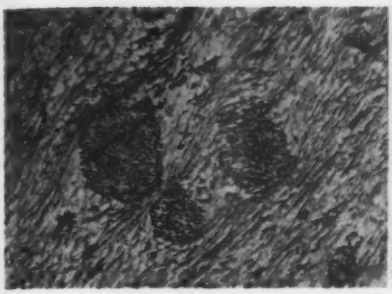


FIGURE 3

FIGURE 4

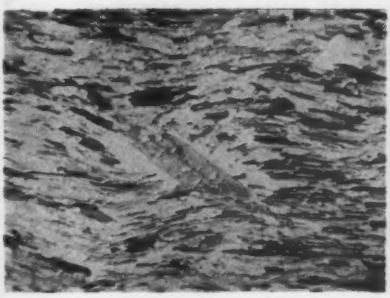


FIGURE 5

FIGURE 6

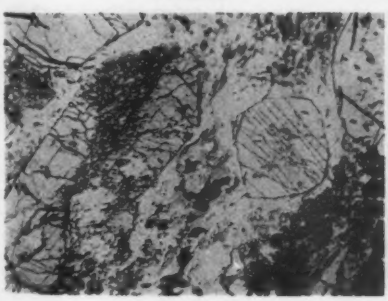


FIGURE 7

FIGURE 8

PHOTOMICROGRAPHS OF TYPICAL SECTIONS OF METAMORPHIC ROCKS FROM THE LEAD DISTRICT

FIGURE 1

FIGURE 2

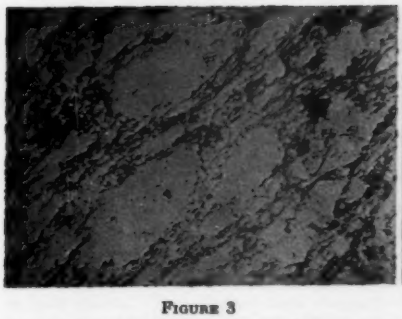


FIGURE 3

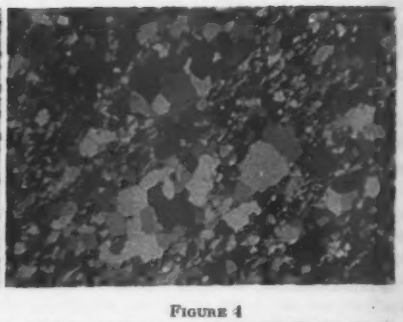


FIGURE 4

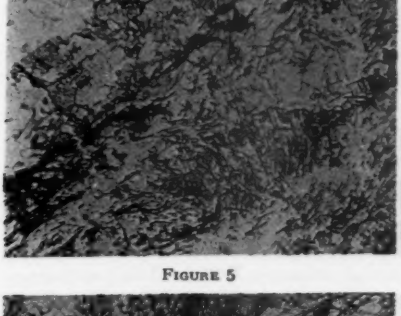


FIGURE 5

FIGURE 6

FIGURE 7

FIGURE 8

PHOTOMICROGRAPHS OF SPECIAL FEATURES IN THE METAMORPHIC ROCKS OF THE LEAD DISTRICT

## PLATE 3.—PHOTOMICROGRAPHS OF SPECIAL FEATURES IN THE METAMORPHIC ROCKS OF THE LEAD DISTRICT

ALL × 20

FIGURE 1. Incipient crystallization of garnet in quartzite. Consists of quartz and a little muscovite, biotite, garnet, tourmaline, and graphite. The muscovite is in random plates and lines of plates. The garnet is in a ragged branching group of tiny grains or a skeletal crystal. The branch to the right extends along a line of tiny muscovite plates. There are two or three tiny random plates of biotite. Plane polarized light.

FIGURE 2. Incipient crystallization of garnet in quartzite. Consists of quartz and a little muscovite, biotite, garnet, tourmaline, and graphite. Quartz is in good dimensional orientation. The muscovite is in tiny well oriented plates. Garnet is in irregular lines of tiny plates parallel to the bedding and the orientation of quartz and muscovite. There are a few random plates of biotite. Plane polarized light.

FIGURE 3. Disappearance of clastic grains of quartz in quartzite. Consists of quartz and a little biotite, muscovite, tourmaline, iron oxide, and graphite. The outlines of the original clastic grains are clearly visible. Plane polarized light.

FIGURE 4. Same as 3, but x-nicols. The grains of clastic quartz have been granulated and begin to merge with the interstitial quartz.

FIGURE 5. Incipient formation of cummingtonite in sideroplesite schist in the outer part of the garnet zone. Consists of sideroplesite, quartz, biotite, muscovite, daphnite, cummingtonite, and a little graphite. Sideroplesite (upper left corner of the field) is in a line of very ragged grains. Rest of the field is a mosaic of tiny quartz grains with small crystals of biotite, muscovite, cummingtonite, and daphnite, all of which replace quartz and to some extent sideroplesite. Plane polarized light.

FIGURE 6. Same as 5, but x-nicols.

FIGURE 7. Replacement of garnet by cummingtonite in inner part of the garnet zone. Consists of cummingtonite, garnet, and a little biotite, carbonate, and graphite. Elsewhere in the section there is a little quartz. Garnet has been replaced by cummingtonite and is veined by carbonate. Plane polarized light.

FIGURE 8. Same as 7, but x-nicols.

42. B
43. M
44. R
45. B
46. E
47. M
48. N
49. O
50. B
51. M
52. G
53. S
54. E
55. O
56. C
57. R
58. O
59. U
60. G
61. G

Occasi
1. Sw
2. P
3. E
4. G
5. T
9. A
10. R
11. M
12. R
13. P
14. B
15. T
16. E
17. P
18. L
19. G
20. C
21. O
22. R
23. T
24. I
25. E
26. B
27. A
28. O
29. B
30. M

## *Publications of The Geological Society of America—Continued*

42. Bibliography of Fossil Vertebrates, 1934–1938. C. L. Camp, D. N. Taylor, and S. P. Welles. 1942. 663 p. $5.70.
43. Miocene Fishes of Southern California. L. R. David. 1943. 193 p., 16 pls., 39 figs. $2.00.
44. Revision of the Suborders, Families, and Genera of the Scleractinia. T. W. Vaughan and J. W. Wells. 1943. 363 p., 51 pls., 39 figs. $3.90.
45. Bibliographic and Faunal Index of Paleozoic Pelmatozoan Echinoderms. R. S. Bassler and M. W. Moodey. 1943. 734 p. $5.70.
46. Evolution and Classification of Paleozoic Crinoids. R. C. Moore and L. R. Laudon. 1943. 153 p., 14 pls., 12 figs. $1.20.
47. Mesozoic and Cenozoic Arcidae from the Pacific Slope of North America. P. W. Reinhart. 1943. 117 p., 15 pls., 3 figs. $1.50.
48. North American Pleospongia. V. J. Okulitch. 1943. 112 p., 18 pls., 19 figs. $1.50.
49. Ozarkian and Canadian Cephalopods. Part II: BreviconES. E. O. Ulrich, A. F. Foerste, and A. K. Miller. 1943. 240 p., 70 pls., 15 figs. $3.00.
50. Bibliography of Economic Geology of South America. J. T. Singewald, Jr. 1943. 159 p. $1.00.
51. Miocene Radiolarian Faunas from Southern California. A. S. Campbell and B. L. Clark. 1944. 76 p. 7 pls. $0.85.
52. Geology and Paleontology of the Permian Area, Northwest of Las Delicias, Southwestern Coahuila, Mexico. R. E. King, C. O. Dunbar, P. E. Cloud, Jr., and A. K. Miller. 1944. 172 p., 45 pls., 20 figs. $2.40.
53. Study and Revision of *Archimedes* (Hall). G. E. Condra and M. K. Elias. 1944. 243 p., 41 pls., 6 figs. $3.00.
54. Early Upper Cambrian Faunas of Central Montana. Christina Lochman and Donald Duncan. 1944. 181 p., 19 pls., 2 figs. $1.80.
55. Origin of Mammals Based upon Cranial Morphology of Therapsid Suborders. E. C. Olson. 1944. 136 p., 27 figs. $1.20.
56. Check List of California Tertiary Marine Mollusca. A. M. Keen and H. Bentson. 1944. 280 p., 4 figs. $2.25.
57. Radiolaria from Upper Cretaceous of Middle California. A. S. Campbell and B. L. Clark. 1944. 61 p., 8 pls., 2 figs. $0.70.
58. Ozarkian and Canadian Cephalopods. Part III: LongiconES. E. O. Ulrich, A. F. Foerste, A. K. Miller, A. G. Unklesbay. 1944. 226 p., 68 pls., 9 figs. $3.00.
59. Upper Devonian Corals of the Mackenzie River Region, Canada. Stanley Smith. 1945. 126 p., 35 pls. $2.00.
60. Glacial Map of North America: Part 1. Glacial map (in color); Part 2. Explanatory notes. By Committee on Glacial Map, R. F. Flint, *Chairman*. 1945. $2.00.
61. Geological Map of South America: Part 1. Preliminary geological map (in black and white); Part 2. Foreword and explanation of legend. By South American Map Committee, A. I. Levorsen, *Chairman*; G. W. Stose, *Editor*. 1945. $2.00.

### MEMOIRS

Occasionally. Continuing orders accepted.

1. Stratigraphy of Western Newfoundland. Charles Schuchert and C. O. Dunbar. 1934. 123 p., 11 pls., 8 figs. $2.40.
2. Paleozoic Plankton of North America. Rudolf Ruedemann. 1934. 141 p., 26 pls., 6 figs. $3.40.
3. Evolution of the Congo Basin. A. C. Veatch. 1935. 183 p., 10 pls., 8 figs. $2.10.
4. Geology in the Coastal Atlas of Western Algeria. R. van V. Anderson. 1936. 450 p., 19 pls., 2 figs. $4.20.
5. The Universal Stage. R. C. Emmons. 1943. 205 p., 13 pls., 95 figs. $2.40.
9. American Tertiary Foraminifera and Corals. T. Wayland Vaughan and John W. Wells. 1945. 200 p., 49 pls., 11 figs. $3.20.
10. Radiolaria from the Kreyenhagen Formation near Los Banos, California. B. L. Clark and A. S. Campbell. 1945. 66 p., 7 pls. $1.20.
11. Mollusca of the Tertiary Formations of Northeastern Mexico. Julia Gardner. 1945. 332 p., 28 pls. 1 fig. $3.90.
12. Revision of the Upper Cambrian Faunas of New Jersey. B. F. Howell. 1945. 46 p., 8 pls. $1.00.
13. Pennsylvanian Geology of a Part of the Southern Appalachian Coal Field. H. R. Wanless. 1946. 163 p., 40 pls., 1 fig. $3.90.
14. Brachiopoda of the Independence Shale. M. A. Stainbrook. 1945. 74 p., 6 pls., 2 figs. $1.20.
15. Tungsten Mineralization in the United States. P. F. Kerr. 1946. 241 p., 23 pls., 34 figs. $3.00.
16. Eocene Faunas from the Department of Bolivar, Colombia. B. L. Clark and J. W. Durham. 1946. 126 p., 28 pls., 1 fig. $2.10.
17. Permian Fusulinids of California. M. L. Thompson, H. E. Wheeler, and J. C. Hazzard. 1946. 77 p., 18 pls., 4 figs. $1.50.
18. Lineation: A Critical Review and Annotated Bibliography. Ernst Cloos. 1946. 122 p., 10 pls., 15 figs. $1.80.
19. Graptolites of North America. Rudolf Ruedemann. 1947. 652 p., 92 pls., 1 fig. $9.00.
20. Corals from the Gulf of California and the North Pacific Coast of America. J. Wyatt Durham. 1947. 68 p., 14 pls., 2 figs. $1.70.
21. Origin and Development of Craters. T. A. Jaggar. 1947. 508 p., 87 pls., 14 figs. $6.00.
22. Reconnaissance Geology of Portions of Victoria Island and Adjacent Regions, Arctic Canada. A. L. Washburn. 1947. 142 p., 32 pls., 4 figs. $4.00.
23. Tertiary Nautiloids of the Americas. A. K. Miller. 1947. 234 p., 100 pls., 30 figs. $4.50.
24. Intertonguing of Marine and Nonmarine Upper Cretaceous Deposits, New Mexico, Arizona, and Southwestern Colorado. W. S. Pike, Jr. 1947. 103 p., 12 pls., 7 figs. $2.25.
25. Evolution of the Horse Brain. Tilly Edinger. 1948. 177 p., 4 pls., 24 figs. $3.00.
26. Bibliographic Index of Permian Invertebrates. C. C. Branson. 1948. 1049 p. $9.50.
27. Assigned but not yet printed.
28. Origin of Granite (Conference). James Gilluly, *Chairman*. 1948. 139 p., 13 pls., 5 figs. $1.80.
29. Batholith and Associated Rocks of Corona, Elsinore, and San Luis Rey Quadrangles, California. E. S. Larsen, Jr. 1948. 5 pls., 16 figs. $4.50.
30. Mineralogical and Structural Evolution of the Metamorphic Rocks. F. J. Turner. 1948. 86 figs. $3.00.

## *Publications of The Geological Society of America*

Publications may be purchased at indicated prices on order accompanied by remittance addressed to The Geological Society of America, 419 West 117 Street, New York 27, N. Y.

### BULLETIN

Monthly. Advance subscription, $10 per year; separate numbers $1.00; back yearly volumes, $12.00 per volume. Ten-year indexes covering vols. 11–20; 21–40; 41–50. $1.50 per copy.

### PROCEEDINGS

Annually. Contains business of Annual Meeting, reports of committees, and memorials. $1.50 per volume.

### HISTORY

THE GEOLOGICAL SOCIETY OF AMERICA, 1888–1930: A Chapter in Earth Science History. By H. L. Fairchild. 1932. 232 p., 18 pls. $2.50.

### BIBLIOGRAPHY OF GEOLOGY EXCLUSIVE OF NORTH AMERICA

Annually, 1933–1940; biennially, 1941–1946. By J. M. Nickles, Marie Siegrist, and Eleanor Tatge. Vols. 1–8 inclusive, $2.50 per volume; Volume 9, $2.75; Volume 10, $1.80; Volume 11, $3.25.

### GEOLOGICAL MAP OF NORTH AMERICA

In color; scale 1:5,000,000. Consists of two sheets each 36 inches by 52 inches. 1946. $3.50.

### SPECIAL PAPERS

Occasionally. Continuing orders accepted.

1. BIBLIOGRAPHIC INDEX OF PALEOZOIC OSTRACODA. R. S. Bassler and Betty Kellett. 1934. 500 p., 24 figs. $3.00.
2. FOSSIL CRUSTACEA OF THE ATLANTIC AND GULF COASTAL PLAIN. M. J. Rathbun. 1935. 160 p., 26 pls., 2 figs. $1.50.
3. FOSSIL NON-MARINE MOLLUSCA OF NORTH AMERICA. Junius Henderson. 1935. 313 p. $2.10.
4. NUCULID BIVALVES OF THE GENUS ACILA. H. G. Schenck. 1936. 149 p., 18 pls., 15 figs. $1.50.
5. GEOLOGY OF SAINT JOHN, NEW BRUNSWICK. A. O. Hayes and B. F. Howell. 1937. 146 p., 9 pls., 4 figs. $1.50.
6. HISTORY OF THE GRAND CANYON OF THE YELLOWSTONE. A. D. Howard. 1937. 159 p., 21 pls., 31 figs. $1.80.
7. ATLANTIC SUBMARINE VALLEYS OF THE UNITED STATES AND THE CONGO SUBMARINE VALLEY. A. C. Veatch and Paul A. Smith. 1939. 101 p., 5 charts, 10 pls., 28 figs. $4.80.
8. MIDDLE DEVONIAN CORALS OF OHIO. G. A. Stewart. 1938. 120 p., 20 pls., 2 figs. $1.50.
9. FOSSIL SNAKES OF NORTH AMERICA. C. W. Gilmore. 1938. 96 p., 4 pls., 38 figs. $0.90.
10. STRATIGRAPHY AND MOLLUSCA OF THE EOCENE OF WESTERN OREGON. F. E. Turner. 1938. 130 p., 22 pls., 7 figs. $1.50.
11. GEOLOGY AND PALEONTOLOGY OF THE MINGAN ISLANDS, QUEBEC. W. H. Twenhofel. 1938. 132 p., 24 pls., 1 fig. $1.50.
12. TERTIARY FLORA FROM THE RIO PICHILEUFU, ARGENTINA. E. W. Berry. 1938. 149 p., 56 pls., 11 figs. $2.10.
13. OZARKIAN AND CANADIAN BRACHIOPODA. E. O. Ulrich and G. A. Cooper. 1938. 323 p., 58 pls., 14 figs. $3.90.
14. DEVONIAN AMMONOIDS OF AMERICA. A. K. Miller. 1938. 262 p., 39 pls., 41 figs. $3.70.
15. CAMBRIAN SYSTEM (RESTRICTED) OF THE SOUTHERN APPALACHIANS. C. E. Resser. 1938. 140 p., 16 pls. $1.50.
16. LOWER CRETACEOUS DEPOSITS IN CALIFORNIA AND OREGON. F. M. Anderson. 1938. 339 p., 84 pls., 2 figs. $4.20.
19. CAMBRIAN MEROSTOMATA. G. O. Raasch. 1939. 146 p., 21 pls., 14 figs. $1.50.
21. PALEONTOLOGY AND STRATIGRAPHY OF THE SILURIAN ROCKS OF THE PORT DANIEL-BLACK CAPE REGION, GASPE. S. A. Northrop. 1939. 302 p., 28 pls., 1 fig. $3.00.
22. LOWER MISSISSIPPIAN ROCKS OF THE EAST-CENTRAL INTERIOR. P. B. Stockdale. 1939. 248 p., 36 pls., 2 figs. $2.40.
23. BIBLIOGRAPHIC INDEX OF NORTH AMERICAN DEVONIAN CEPHALOPODA. E. M. Kindle and A. K. Miller. 1939. 179 p. $1.20.
25. DEVONIAN STRATIGRAPHY AND PALEONTOLOGY OF THE ROBERTS MOUNTAINS REGION, NEVADA. C. W. Merriam. 1940. 114 p., 16 pls., 7 figs. $1.50.
26. PERMIAN AMMONOIDS OF THE GUADALUPE MOUNTAIN REGION AND ADJACENT AREAS. A. K. Miller and W. M. Furnish. 1940. 242 p., 44 pls., 59 figs. $2.70.
27. BIBLIOGRAPHY OF FOSSIL VERTEBRATES 1928–1933. C. L. Camp and V. L. VanderHoof. 1940. 503 p. $4.20.
28. REVIEW OF THE PELYCOSAURIA. A. S. Romer and L. I. Price. 1940. 538 p., 46 pls., 71 figs. $4.20.
29. PHACOPID TRILOBITES OF NORTH AMERICA. D. M. Delo. 1940. 135 p., 13 pls. $1.20.
31. SUBMARINE TOPOGRAPHY OFF THE CALIFORNIA COAST: CANYONS AND TECTONIC INTERPRETATION. F. P. Shepard and K. O. Emery. 1941. 171 p., 18 pls., 42 figs., 4 charts. $3.90.
35. GEOLOGY OF THE "MARTIC OVERTHRUST" AND THE GLENARM SERIES IN PENNSYLVANIA AND MARYLAND. Ernst Cloos and Anna Hietanen. 1941. 207 p., 28 pls., 58 figs. $4.20.
36. HANDBOOK OF PHYSICAL CONSTANTS. Edited by Francis Birch, *Chairman, et al.* 1942. 325 p., 19 figs. $1.70.
37. OZARKIAN AND CANADIAN CEPHALOPODS, PART I, NAUTILICONES. E. O. Ulrich, Aug. F. Foerste, A. K. Miller, and W. M. Furnish. 1942. 157 p., 57 pls., 23 figs. $2.40.
38. TEREBRATULOID BRACHIOPODA OF THE SILURIAN AND DEVONIAN. P. E. Cloud, Jr. 1942. 132 p., 26 pls., 17 figs. $1.80.
39. EOCENE RADIOLARIAN FAUNAS FROM THE MT. DIABLO AREA, CALIFORNIA. B. L. Clark and A. S. Campbell. 1942. 112 p., 9 pls., 5 figs. $1.00.
40. HADROSAURIAN DINOSAURS OF NORTH AMERICA. R. S. Lull and N. E. Wright. 1942. 242 p., 31 pls., 90 figs. $2.70.

*(Continued on inside back cover)*

ESCANEADO